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NORTHERN HYDROLOGY: CANADIAN PERSPECTIVES

Editors

T.D. Prowse and C.S.L. Ommanney

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FOREWORD

Scientists of the National Hydrology Research Institute should, from time-to-time, produce reports which can serve as tools or reference material for scientists, engineers and planners working on environmental problems. This report is the first in a series directed to this function. The editors were able to develop this comprehensive summary of northern hydrological science using contributions solicited from various experts, at the National Hydrology Research Institute and elsewehere. The chapters cover a wide range of hydrological phenomena, including material on water quality and management of water resources. As part of each review, the contributors have indicated where knowledge remains deficient and where research efforts and resources could most profitably be applied. Adequate knowledge of northern hydrological processes and how they may be affected by industrial and muncipal growth will be vital to those responsible for the orderly and sustainable development of the North. Furthermore, with global change likely to affect the historical norms for such things as water supply, the aquatic and permafrost environments, it becomes even more important to understand the environmental factors that influence these processes. This useful and valuable compendium of northern hydrological science will help provide this knowledge.

Science Report No.1, by addressing the scientific issues of northern hydrology, provides authoritative information for reference and for application to northern water-resource problems in Canada.

T.M. Dick, Director

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11 June 1990

PREFACE

The idea for this book originated in 1986 when one of the editors was appointed Head of the Northern Hydrology Section (and more recently the expanded Northern Environment Program), at the National Hydrology Research Institute (NHRI). Although having experience in specific northern hydrological topics, an ability to assess all aspects of the subject was considered essential to building and maintaining a strong team capable of addressing the most important research issues. To help develop this ability, it was decided to have a select group of Canadian experts review the past, present and future of the major fields of northern hydrology. These reviews could then be used to help identify the major research trends and most important knowledge gaps in northern hydrology.

Although a myriad of small topics could have been identified for review, it was decided to focus on broad subject areas; ones considered particularly germane or of special significance to the hydrology of northern latitudes. These were: snow hydrology, permafrost hydrology, ground-water hydrology, hydrology of floating ice, glacier hydrology, regional energy balances, water quality, regional hydrology and water legislation.

As the project evolved, it was decided to release the reviews in book form for the benefit of other northern water scientists, managers and practitioners. To develop consistent coverage among reviews, each author was asked to adhere to the standardized format shown below, although this was difficult for reports not specifically reviewing research (i.e., regional hydrology and water legislation). The editors have re-arranged some original material and inserted text to further maximize consistency and to better link the various chapters.

As is with most endeavours of this type, it required much more effort than originally expected. Timely completion of the book would not have been possible without the superb drafting, graphic design and publication management provided by Philip K. Gregory (NHRI), supported by Malcolm Conly, John Gibson and other NHRI staff. The editors would also like to acknowledge the dedication of Debbie McKnight and the NHRI library staff for their perseverence in unearthing even the most obscure references on northern hydrology.

We hope this book will lead to a greater interest in, and knowledge of, northern science, and help promote an integration of water-related research with northern environmental studies; a goal that is rapidly being recognized as a prerequisite to the solution of many of the evolving problems in the North.

Specific Points to be Covered

1. Introduction
- basic hydrological processes
- variations in the hydrological regime, i.e. geographical extent and range of influence,
- relationships to other northern hydrological regimes
- importance to the northern environment

2. Historical Background
- major historical research thrusts
- major research groups
- location of major research areas
- rationale for such research in the North

3. Current Research
- current research activities and rationale

4. Future Research Needs
- major unknowns in existing research approaches
- new research approaches
- major incentive for future research
- classification of research into: field research, laboratory research, computer/physical/theoretical modelling, etc.
- major stumbling-blocks/impediments to conducting research

General Points
- focus north of 60°N where required to limit discussion
- deal with existing data networks and sources of information, instrumentation, remote sensing, etc.
- reference relevant work only
  (avoid exhaustive literature review)

Summary list of guidelines supplied to chapter authors.

T.D. Prowse and C.S.L. Ommanney, editors
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TABLE OF CONTENTS

FOREWORD i

PREFACE iii

LIST OF FIGURES xiii

LIST OF TABLES xv

1. NORTHERN HYDROLOGY: AN OVERVIEW

1.1. Introduction 1
1.2. Definition of "North" 1
  1.2.1. Physical North 2
    1.2.1.1. Polar Circles 2
    1.2.1.2. Temperature 2
    1.2.1.3. Snow/Ice Covers 3
    1.2.1.4. Frozen Ground 4
    1.2.1.5. Vegetation 5
    1.2.1.6. Run-Off Direction 5
    1.2.1.7. Composite Definition 5
    1.2.1.8. Guideline Definition 6
1.3. Natural Landscape, Climate and Hydrology 7
  1.3.1. General Landscape 7
  1.3.2. Hydroclimatic Characteristics 9
    1.3.2.1. Precipitation/Snowcover 9
    1.3.2.2. Drainage/Run-Off 13
    1.3.2.3. Surface Storage 15
    1.3.2.4. Permafrost 15
    1.3.2.5. Ground Water 16
    1.3.2.6. Evaporation/Evapotranspiration 17
    1.3.2.7. Freshwater Ice Cover 19
    1.3.2.8. Data-Collection Networks 23
1.4. Hydrological Classifications 25
  1.4.1. Church (1974) Classification 25
    1.4.1.1. Arctic-Nival 26
    1.4.1.2. Subarctic-Nival 26
    1.4.1.3. Proglacial 26
    1.4.1.4. Muskeg 26

    1.4.2. Network Classification 27
1.5. Development and Water Use 28
    1.5.1. Historical Resource Development 28
    1.5.2. Present Resources 29
        1.5.2.1. Hydro-Power 29
        1.5.2.2. Diversions 31
        1.5.2.3. Mining 31
        1.5.2.4. Oil and Gas 32
1.6. Evolving Issues 32
    1.6.1. Northern Pollution 32
    1.6.2. Global Change Effects 33
        1.6.2.1. Precipitation/Snow/Glaciers 33
        1.6.2.2. Floating Ice 34
        1.6.2.3. Permafrost 34
        1.6.2.4. Other 35
    1.6.3. Monitoring Global Change 35
1.7. Reflections 35

2. SNOW HYDROLOGY

2.1. Introduction 37
    2.1.1. Basic Processes 38
        2.1.1.1. Atmospheric Processes 38
        2.1.1.2. Snowcover Formation 39
        2.1.1.3. Snowmelt Run-Off Processes 40
        2.1.1.4. Snowcover Chemistry 48
    2.1.2. Geographical Variations 50
        2.1.2.1. Snowfall 50
        2.1.2.2. Snowcover 51
        2.1.2.3. Nival Streamflow Regimes 51
    2.1.3. Relationship to Other Hydrological Variables 52
2.2. Historical Background 52
    2.2.1. Snowfall 53
    2.2.2. Snowcover Properties 54
    2.2.3. Snowmelt Run-Off 55
2.3. Current Research 57
    2.3.1. Blowing Snow 57
    2.3.2. Meltwater Infiltration 57

2.3.3. Water Flux Through Cold Snowcovers 57
2.3.4. Remote Sensing 58
2.3.5. Snow Chemistry 58
2.3.6. Snow/Ground Interface Temperature 58
2.3.7. Snow Radiative Properties 58
2.4. Future Research Needs 58
2.4.1. Snowfall 59
2.4.2. Snowcover 60
2.4.3. Snowmelt 60
2.4.4. Snowmelt Run-Off 60
2.4.5. Removal of Solutes from Melting Snowcovers 61

3. PERMAFROST HYDROLOGY

3.1. Introduction 63
3.2. Historical Background 64
3.2.1. Streamflow Studies 65
3.2.2. Water-Balance Studies 68
3.2.3. Hillslope Hydrological Studies 70
3.2.4. Studies of Lake and Wetland Hydrology 71
3.3. Recent Trends 72
3.4. Current Research 72
3.4.1. Coupling Snowmelt, Infiltration and Soil Moisture 73
3.4.2. Hydrological Modelling 73
3.5. Future Research Needs 75
3.5.1. Regional Hydrology 75
3.5.2. Climatic Change and Permafrost Hydrology 75
3.5.2.1. Application of Spatial Analogues 76
3.5.2.2. Application of Climatic Variability Analogues 76
3.5.2.3. Use of Simulation Models 76

4. GROUND-WATER HYDROLOGY

4.1. Introduction 77
4.1.1. Permafrost 77
4.1.2. Properties of Frozen Ground and Permafrost 77
4.1.2.1. Ice and Water Contents 77
4.1.2.2. Hydraulic Conductivity 79
4.1.3. Effects on Water Chemistry 79

4.1.3.1. Low Temperatures 79
4.1.3.2. Permafrost 79

4.2. Aquifers in Permafrost Areas 80
4.2.1. Position Relative to Permafrost 80
4.2.2. Suprapermafrost Aquifers 81
4.2.2.1. Subtype I Aquifers 82
4.2.2.2. Subtype II and III Aquifers 82
4.2.2.3. Chemical composition of suprapermafrost water 82
4.2.3. Intrapermafrost Aquifers 82
4.2.3.1. Open Taliks 83
4.2.3.2. Lateral Taliks 83
4.2.3.3. Isolated Taliks 83
4.2.3.4. Chemical Composition of Intrapermafrost Water 83
4.2.4. Subpermafrost Aquifers 84
4.2.4.1. Unconsolidated Deposits 84
4.2.4.2. Sedimentary Bedrock 84
4.2.4.3. Karst 85
4.2.4.4. Igneous and Metamorphic Terrain 85
4.2.4.5. Chemical Composition of Subpermafrost Water 85
4.3. Ground-Water Hydrology 87
4.3.1. Comparison with Non-Permafrost Areas 87
4.3.2. Infiltration and Ground-Water Recharge 87
4.3.3. Lateral Movement of Ground Water 88
4.3.4. Discharge of Ground Water 88
4.3.4.1. Springs 88
4.3.4.2. Base Flow 90
4.3.4.3. Icings 91
4.3.4.4. Ground-Water Discharge Rates from Icing Volumes 93
4.3.4.5. Chemical Composition of Icings and Meltwater 93
4.3.5. Importance of Ground Water 94
4.3.5.1. Fisheries 94
4.3.5.2. Vegetation 94
4.3.5.3. Water Supply 94

4.3.5.4. Recreation and Space Heating 95

4.4. Historical Background 95

4.4.1. Pre-1970 Situation 95

4.4.2. Recent Advances 96

4.4.3. Agencies Involved in Recent Research 97

4.4.4. Incentives for Recent Research 98

4.5. Current Research 98

4.6. Future Research Needs 99

4.6.1. Major Unknowns 99

4.6.2. Subjects Requiring Research 99

4.6.3. Future Incentives for Ground-Water Research 100

4.6.4. Major Impediments 101

5. HYDROLOGY OF FLOATING ICE

5.1. Introduction 103

5.2. Ice Processes on Rivers and Lakes: an Overview 108

5.3. Historical Background 114

5.3.1. Floods and High Stages 114

5.3.2. Low Flow 118

5.3.3. Flow Duration 119

5.3.4. Flow Measurement 119

5.4. Current Research 123

5.4.1. Ice Jams, Break-Up and Freeze-Up 123

5.4.2. Low Flows 126

5.4.3. Flow Duration 127

5.4.4. Flow Measurement 127

5.5. Future Research Needs 127

5.5.1. Ice Jams, Break-Up and Freeze-Up 128

5.5.2. Low Flows 130

5.5.3. Continuous Simulation 130

5.5.4. Flow Measurement 130

5.6. New Research Approaches 131

5.7. Conclusion 134

5.8. Acknowledgements 134

6. GLACIER HYDROLOGY

6.1. Introduction 135
6.2. Glacier Hydrology: an Overview 135
- 6.2.1. Run-Off 135
- 6.2.2. Local Variability in Climatic Inputs 136
- 6.2.3. Regional Variability 136
- 6.2.4. Temporal Variability in Climatic Inputs 136
- 6.2.5. Glacier Outburst Floods 136
- 6.2.6. Distribution and Characteristics of Canadian Glaciers 137

6.3. Process and Regional Examples 137
- 6.3.1. Typical Run-off Components from a Glacierized Catchment 138
- 6.3.2. Western Cordilleran Example - Peyto Glacier 142
  - 6.3.3.1. Basin Description 142
  - 6.3.3.2. Meteorological Inputs and Snow-Line Movement 143
  - 6.3.3.3. Internal Hydrology of the Glacier 147
  - 6.3.3.4. Specific Yield within the Basin 147
  - 6.3.3.5. The Resultant Hydrographs 147
  - 6.3.3.6. Detailed Interrelationship (1970/74) 147
- 6.3.4. High-Arctic Example - White Glacier 149

6.4. A History of Glacier Hydrology Studies 150
- 6.4.1. Studies in the Western Cordillera 150
- 6.4.2. The Arctic Islands 152
- 6.4.3. Studies in the Yukon 152
- 6.4.4. Studies in Labrador 152
- 6.4.5. The International Hydrological Decade 152
- 6.4.6. International Hydrological Programme 155

6.5. The Main Achievements of Hydrology Studies in Glacierized Basins 156
- 6.5.1. Basic Description of Resources 156
- 6.5.2. Achievements in Glacier Mass Balance 156
- 6.5.3. Achievements in Glacier-Run-Off Studies 159
  - 6.5.3.1. Meltwater Generation 159
  - 6.5.3.2. Glacier Outburst Floods 160
  - 6.5.3.3. Models of Basin Run-Off 160
- 6.5.4. Achievements in Surveying and Remote Sensing 160

6.6. The Future 161

7. WATER-QUALITY RESEARCH

7.1. Introduction 163
7.2. Historical Background 164
- 7.2.1. Activities Prior to 1969 164
- 7.2.2. Overview of N.W.T. Water-Quality Network Data 165
- 7.2.3. Mackenzie River Basin 171
- 7.2.4. Yukon River Basin 174
- 7.2.5. Limnological Research Activities near Resolute, Cornwallis Island and at Saqvaqjuac 176

7.3. Current Research - Long-Range Atmospheric Transport and Deposition 178
7.4. Future Research Needs 184
7.5. Acknowledgements 186

8. THE REGIONAL ENERGY BALANCE

8.1. Introduction 187
- 8.1.1. Basic Processes 188
- 8.1.2. Regional Variations 189
- 8.1.3. Relationship to Other Hydrological Variables 191
- 8.1.4. Importance to the Northern Environment 193

8.2. Historical Background 193
- 8.2.1. Subarctic 193
- 8.2.2. Arctic Tundra 195
- 8.2.3. Glacierized Basins 200

8.3. Current Research 204
8.4. Future Research Needs 205

9. REGIONAL HYDROLOGY

9.1. Introduction 207
9.2. Hydrological Regions 208
- 9.2.1. Shawinigan Engineering Studies 208
- 9.2.2 Acres Consulting Services Ltd. 210
  - 9.2.2.1. Annual Yield 210
  - 9.2.2.2. Low Flow 211
  - 9.2.2.3. Floods 211
- 9.2.3. Joint DOE-DIAND Studies 211
- 9.2.4. 1985-86 IWD Network Evaluation in Western and Northern Region 212

9.2.5 The Mackenzie River Basin Flow Model 217
9.3. River Elements in the Northern Hydrological Cycle 218
9.3.1. Snowfall Components 220
9.3.2. Ice Processes and Northern Streamflow 221
9.3.2.1. The Freeze-Up 221
9.3.2.2. Midwinter Flows 221
9.3.2.3. Streamflow in the Annual Break-Up 222
9.4. Northern Streamflow Data Collection 223
9.5. Conclusions 225

10. WATER MANAGEMENT NORTH OF 60°

10.1. Introduction 227
10.2. Legislative Framework 227
10.3. Water Boards 228
10.4. Licensing and Authorization 228
10.4.1. Licensing 228
10.4.2. Authorization 229
10.5. Enforcement 230
10.6. Data Collection 230
10.6.1. Water Quantity 230
10.6.2. Water Quality 230
10.6.3. Precipitation 231
10.7. Planning 231
10.7.1. Comprehensive Planning 231
10.7.2. Integrated Planning 234
10.8. Major Management Concerns 236
10.8.1. Mine Abandonment 236
10.8.2. Transboundary Issues 236
10.9. Need for Legislative Amendments 237
10.10. Aboriginal Land Claims 237
10.11. Transfer of Responsibilities to Territorial Governments 239

11. BIBLIOGRAPHY 241

APPENDIX A - List of Symbols 303

APPENDIX B - List of Abbreviations and Acronyms 307

LIST OF FIGURES

1.1. Limits of the Northern Regions 3
1.2. Major Drainage Basins and Annual River Discharge 6
1.3. Physiographic Regions of Canada 7
1.4. Transboundary Flow of the Mackenzie River 8
1.5. Mean Annual Total Precipitation 10
1.6. Mean Annual Total Rainfall 10
1.7. Mean Annual Total Snowfall..........11
1.8. Dates of Formation and Loss of Snowcover 12
1.9. Mean Maximum Depth of Snow 12
1.10. Wetland Regions of Northern Canada 13
1.11. Permafrost Distribution 14
1.12. Mean Annual Lake Evaporation 18
1.13. Derived Evapotranspiration 18
1.14. Mean Maximum Ice Thickness in Lakes 20
1.15. Mean Maximum Ice Thickness in Rivers 20
1.16. Mean Freeze-Over Date on Lakes 21
1.17. Mean Freeze-Over Date on Rivers 21
1.18. Mean Ice-Free Date on Lakes 22
1.19. Mean Ice-Free Date on Rivers 22
1.20. Number of Hydrometric Stations by Basin Area..........24
1.21. Length of Record of Hydrometric Stations..........24
1.22. Selected Development Opportunities in the Mackenzie River Basin..........30
2.1. Flow Fingers in a Layered Snowcover..........42
2.2. Water Movement into Cold Dry Snow..........43
2.3. Predicted Snow-Cover Hydrographs 46
2.4. Relationship Between Snow Density, Grain Size and Saturated Permeability..........47
2.5. Lag Times for Vertical Unsaturated and Horizontal Saturated Flow..........49
3.1. Hydrographs of Typical Streamflow Regimes in Permafrost Areas..........66
3.2. Typical Arctic-Nival Regime Run-Off, McMaster River..........67
3.3. Water Balances for McMaster River Basin..........69
3.4. Overland Flow During the Snowmelt Period..........71
3.5. Flow Chart for the Modelling of Hydrological and Thermal Processes 74

4.1. Variation of Temperature with Depth in Permafrost 78
4.2. Unfrozen Water Content vs. Ground Temperature 78
4.3. Aquifers in Permafrost Areas 81
4.4. Artesian Discharge from Talik Below a Pingo 84
4.5. Large Sinkhole Depression in Collapsed Karst 86
4.6. Open Water Maintained by Ground-Water Discharge 88
4.7. Layered Icing and Development of a Thermo-Erosional Niche 89
4.8. Frost Mound at the Bear Rock Discharge Area 89
4.9. Daily Discharge at Snag Creek Gauging Station 90
4.10. Spring- and Creek-Water Chemistry, Vermilion Creek 91
4.11. Chemical Composition of Ogilvie River 92
5.1. Ice-Jam Flood, Fredericton, New Brunswick, 1936 104
5.2. Exceedence Probability for Ice Jams and Summer Floods ... 104
5.3. Winter Hydrograph of the Clearwater River at Draper 105
5.4. Minimum Flow for Streams in Northern Alberta 105
5.5. Mackenzie Delta Break-Up and Ice Jam Flooding 106
5.6. Ice Jam Flood, St. Lawrence River at Montréal 107
5.7. Outlet of Lake Laberge, Yukon 110
5.8. Lodgement, North Saskatchewan River near Myrnam 110
5.9. Loss of Discharge to Channel Storage During Ice-Cover Progression 112
5.10. Ice Jam and Break-Up Ice Jam, Hay River 115
5.11. Water Temperature Beneath an Ice Jam, Liard River 116
5.12. Typical Small Stream in Winter 118
5.13. Dependence of Errors in Interpolated Winter Discharge 120
5.14. Open Water Rating Curve and Winter Discharge Measurements: Takhini River 121
5.15. Baillie River Spring Hydrograph and Ice Conditions 122
5.16. Pre-Break-Up Open Water, Athabasca River 124
5.17. Freeze-Up Pack on the Exploits River 126
5.18. Downward-Looking Sonar for Measuring Water Levels 128
6.1. Glacier Distribution, Arctic Islands 138
6.2. Glacier Distribution, Western Canada 139
6.3. Hydrograph for Partially Glacierized Catchment 140
6.4. Percentage of Mean Monthly to Annual Discharge 140
6.5. Hydrographs for Two Streams, Rocky Mountains 141
6.6. Peyto Glacier Basin; Elevation and Terrain Types 142

6.7. Peyto Glacier Basin; Total Global Radiation 145
6.8. Peyto Glacier Basin; Net Short-Wave Radiation 145
6.9. Peyto Glacier; Variation in Snow-Line Position.....................146
6.10. Maps of Specific Annual Yield for 1970 and 1974................146
6.11. Hydrographs for Peyto Creek, 1967-1974 148
6.12. Snow-Line Movement, Meteorology and Discharge..............149
7.1. Water-Quality Network Stations, Resolute and Saqvaqjuac..166
7.2. Location of Snow-Sampling Sites, 1986 and 1987 180
7.3. Trend of Total PCB Residues in Agassiz Ice Cap..................184
8.1. Latitudinal Variation in Radiation Terms at the Surface........190
9.1. Yukon and N.W.T. Hydrological Regions 213
9.2. N.W.T. Total Run-Off for 1985...214
9.3. Central Arctic Regions and 1985 Annual Yield......................215

LIST OF TABLES

Summary List of Guidelines Supplied to Chapter Authors iv
1.1 Cold Regions 4
4.1. Physical and Chemical Parameters for Ground Waters 80
4.2. Ground-Water Use North of 60°N 95
6.1. Peyto Glacier Mass Balance..143
6.2. Peyto Glacier Meteorological Elements and Snow-Line Movement 144
6.3. Peyto Glacier Derivation of Summer Streamflow..................149
6.4. Glacier Mass Balance and Hydrological Records..................151
6.5. Locations of the Canadian Glacier Basins 154
6.6. Physical Characteristics of the Glacier Basins 154
6.7. Basic Data Collections and Major Agencies Involved...........154
7.1 Selected Water-Quality Data for N.W.T. Stations 167
7.2 Organochlorine Content of Arctic Snowpacks 180
7.3 Contaminant Deposition in Precipitation and Snow 181
7.4 Comparion of Hexacholorocyclohexane Concentrations.......182
8.1. Surface Energy Balance Studies in the Subarctic 194
8.2. Surface Energy Balance Studies on Arctic Tundra 197
8.3. Surface Energy Balance Studies in Arctic Glacierized Basins 201
8.4. Summary of the Surface Energy-Balance Studies..................203
9.1. Acres (1976) Classification of Hydrological Regions 210
9.2. Hydrological Regions of the Northwest Territories 213

9.3. Ratios of Maximum to Minimum Monthly Flows 218
9.4. Stations in the Yukon and N.W.T. with no Backwater from Ice 223
10.1. Mackenzie River Basin Study: Recommendations 233
10.2. Mackenzie River Basin Study: List of Supplemental Reports 234
10.3. Yukon River Basin Study: Recommendations 235
10.4. Northern Inland Waters Act: Major Proposed Amendments 238

1

NORTHERN HYDROLOGY: AN OVERVIEW

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1.1. INTRODUCTION

Since many readers of this book probably do not have a general background in northern hydrology, an overview of the North's unique character and problems is provided in the following sections of this chapter. Cross-references are made to other chapters for more detailed treatments of specific subjects. To form a geographical frame of reference for "northern" hydrology, Section 1.2. reviews the various definitions of the "North", both in a historical and physical sense. The major geographical regions of the North are examined next, focusing on the basic landforms, climatology and hydrology. Section 1.4. describes the major hydrological classifications of the North, and 1.5. reviews major uses of water and developments that may place further demands on the hydrological system. Two evolving issues, northern pollution and global change, are discussed in Section 1.6. and some brief comments for reflection conclude the chapter.

1.2. DEFINITION OF "NORTH"

Webster's defines "northern" simply as "in, of, toward, or facing the north" (Guralnik, 1986). Hence, any work conducted north of the equator could be considered "northern". Moving northward, the term should become easier to define but the reverse appears to be true. To make matters worse, the perception of what is "North" has changed with time, largely because of the "taming" effects produced by advancing development. Immediately following Confederation, for example, when even the western provinces were Canada's frontier, anything beyond Lake Nipissing was termed "North" (Zaslow, 1971, p.xi). As development expanded westward and northward so the northern boundary migrated poleward. Unfortunately, however, Canada's population became concentrated in a southern fringe along the United States' border and the "North" remains largely unknown to a majority of Canadians today. As such, it is the source of many visions including: "... The southern vision (the North is a hinterland to be exploited for the benefit of southern Canada)... The romanticized vision (wilderness must never be touched)... The pessimistic vision (which sees only the problems) and the developmental vision (with natural gas opportunities)" (W.P. Wilder, quoted by Hamelin, 1979, p.6). Even northern inhabitants hold different visions of the "North" and "... have great difficulty picturing the North in its geographic immensity; theirs is still a fragmented North." (Hamelin, 1979, p.7).

1.2.1. Physical North

One might think that using physical boundaries to define the "North" would avoid the pitfalls of the transient nature of social definitions. They too, however, are not consistent and, although largely constant with time, are often spatially contradictory.

1.2.1.1. Polar Circles

A restrictive but frequently-used definition of the physical North is the area above the Arctic Circle (Figure 1.1); the 66°33'N line of latitude above which the sun does not rise for one or more days during the winter and does not set for one or more days in summer. A common misconception is that the geographical term "Arctic" refers to this region. Although it is defined on the basis of the radiation regime, the actual Arctic is defined according to an air temperature regime as described below.

1.2.1.2. Temperature

Air temperature is perhaps the variable most commonly used to characterize the cold, northern regions. Köppen's (1936) classification considered this to include: a mean temperature of the coldest month not exceeding -3°C and no more than 4 months with a mean temperature greater than 10°C.

The common terms "Arctic" and "Subarctic" are also based simply on coldness criteria (e.g., Washburn, 1953). The arctic classification (Figure 1.1) refers to those regions where the mean temperature for the warmest month remains below 10°C and the mean annual temperature below the freezing point. Further subdivision into "High" and "Low" Arctic (Figure 1.1) is achieved by considering July temperatures. The former applies when the mean July temperature is below 5°C and the latter when it falls between 5° and 10°C. The Subarctic refers to those regions where the mean temperature for the coldest month is below freezing and the mean temperature of the warmest month is above 10°C but, as with the Köppen classification, no more than 4 months have a mean temperature exceeding 10°C.

Perhaps the broadest and simplest coldness definition of the northern cold regions is that by Gerdel (1969) who used the 0°C mean annual isotherm of the coldest month as the southern limit (Figure 1.1). This, however, results in a North extending as far south as Sante Fe, New Mexico (latitude 36°N).

A myriad of other air-temperature combinations could and have been used to define the North, but problems of representativeness arise from the use of such simple static definitions. For example, how representative is the mean January temperature when the same isotherm that characterizes the southern portion of the Yukon also runs through portions of the mid-western United States; or the mean July temperature, when the summer temperatures are warmer in Yellowknife than in St. John's, Newfoundland?

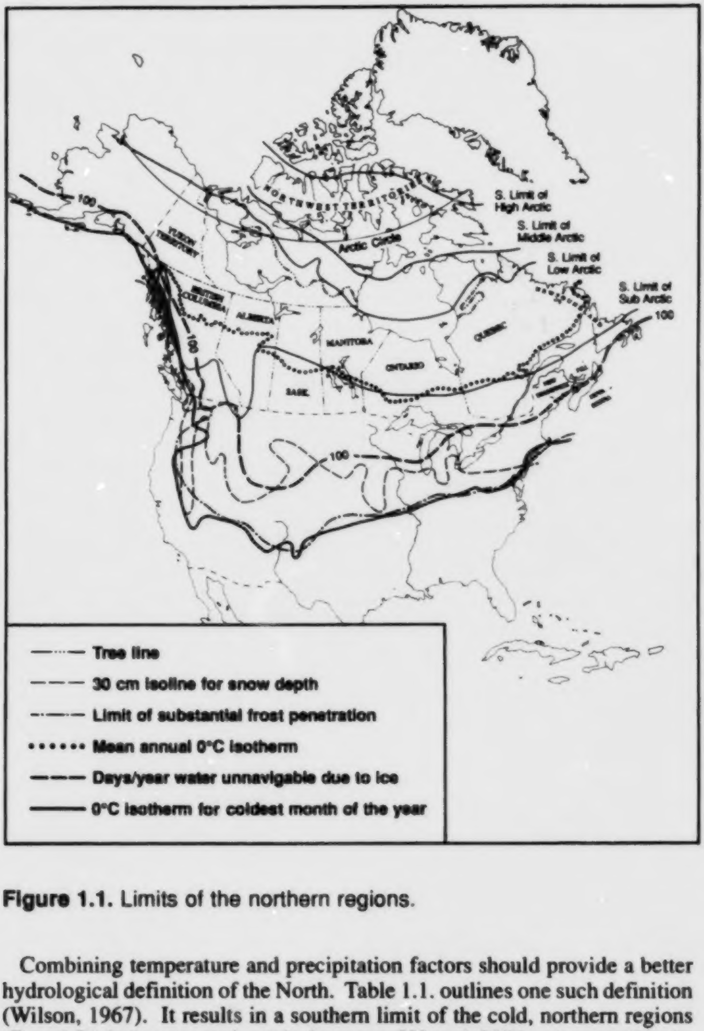

S. Limit of High Arctic
S. Limit of Middle Arctic
S. Limit of Low Arctic
S. Limit of Sub Arctic
100
Arctic Circle
NORTHWEST TERRITORIES
YUKON TERRITORY
BRITISH COLUMBIA
ALBERTA
SASK.
MANITOBA
ONTARIO
QUEBEC
Tree line
30 cm isoline for snow depth
Limit of substantial frost penetration
Mean annual 0°C isotherm
Days/year water unnavigable due to ice
0°C isotherm for coldest month of the year

Figure 1.1. Limits of the northern regions.

Combining temperature and precipitation factors should provide a better hydrological definition of the North. Table 1.1. outlines one such definition (Wilson, 1967). It results in a southern limit of the cold, northern regions (E and D) being approximately between 50° and 55°N. Apparently, the precipitation factor is not that significant since the boundary closely parallels that of the southern limit of the Subarctic, a simple temperature-based definition (Figure 1.1).

1.2.1.3. Snow/Ice Covers

Most would agree that typical "northern" hydrological conditions include the presence of snow and ice covers. However, even if conservative estimates of the latitudinal extent of these parameters are used, they encompass much of North America. For example, freshwater systems are unnavigable because of ice conditions for 100 days per year as far south as 42°N and, even excluding mountainous regions, snowfall regularly accumulates to more than 30 cm as far south as 38°N (Bates and Bilello, 1966; Figure 1.1).

1.2.1.4. Frozen Ground

Permafrost, or perennially-frozen ground, is another integral part of the northern hydrological environment. The boundaries for continuous and discontinuous permafrost closely follow those for the arctic and subarctic climates (Figure 1.1). This is not surprising, however, since the lack of detailed field information has necessitated the more or less arbitrary drawing of permafrost boundaries according to air-temperature isotherms (see Section 1.3.2.4.).

E	**Polar (snow) Climate:** mean temperatures of warmest month below 10°C.		ET	**Tundra:** mean temperature of warmest month above 0°C but below 10°C.
			EF	**Frost:** mean temperature of every month below freezing, permanent snow.
			E	**Falkland:** mean temperature of coldest month greater than -3°C (26.6°F), but warmest month below 10°C.
			EH	**Mountain:** mountain type of lower latitudes (greater than 1500 m).
D	**Cold, snow-forest Climate:** mean temperature for coldest month less than -3°C, warmest month above 10°C.	seasonal precipitation distribution	f	no dry season, precipitation of wettest month in cold season less than three times that of driest month in warm season.
			w	dry winters, driest month of cold season less than 1/10 precipitation of wettest month of warm season.
		severity of winter temperature	c	mean temperature of coldest month between -38°C (-36.4°F) and -3°C; one to four months with mean temperature between 10°C and 22°C (71.6°F).
			d	mean temperature of coldest month less than -38°C; one to four months with mean temperature between 10°C and 22°C.

Table 1.1. Cold regions *(after Wilson 1967)*.

In an attempt to more completely encompass the northern areas affected by frozen ground, Bates and Bilello (1966) identified a southern limit of substantial frost penetration (approximately 0.3 m [1 ft] once every 10 years; Figure 1.1). As for the permafrost boundaries, however, this line was estimated, not from ground-based information, but from air-temperature data, specifically that of 56 accumulated Celsius freezing degree-days (100 accumulated Fahrenheit freezing degree-days).

1.2.1.5. Vegetation

Tree line (Figure 1.1) is one biological factor sometimes used to define the North and is useful in a hydrological sense because it reflects the vegetative effects produced by both temperature and precipitation. Although tree line crudely approximates the 10°C mean July isotherm, it does not form a precise boundary and is often only a broad transition zone, complicated further by altitudinal effects. Similar problems exist in the use of related vegetation boundaries, such as the tundra-taiga transition which in the Quebec-Labrador peninsula corresponds to a belt several hundred kilometres wide and could be considered a separate zone in itself (Hamelin, 1979).

1.2.1.6. Run-Off Direction

From the point of view of run-off, the areal extent of northward-flowing basins is immense (Figure 1.2). 84% of the Canadian land mass is drained by rivers that flow northward into either the Arctic Ocean, Bering Sea or Hudson Bay (after Greene, 1978). Obvious problems arise, however, when one attempts to consider as "northern", some northward-flowing rivers that have headwaters in the mid-western United States, such as the South Saskatchewan and Red Rivers.

1.2.1.7. Composite Definition

Clearly difficulties exist when attempts are made to define "North" based on a single physical characteristic. Hamelin (1979) found that no less than 20 separate criteria have been employed to define high-latitude cold regions and have resulted in an equal number of separate boundaries.

In the most rigorous effort to date, Hamelin (1979) attempted to obtain one comprehensive boundary using a suite of ten, converging factors, many of which are significant to the hydrological cycle. The list included: latitude; summer heat above 5.6°C; annual cold below 0°C; types of ice (in the ground, on land, or on water); total precipitation; development of the vegetation cover; accessibility (air or surface); number of inhabitants or regional population density; and finally, degree of economic activity. Largely because of the latter criteria, "nordicity" varies temporally. During the last century, Hamelin (1979) estimates there has been approximately a 25% denordification of Canada's high-latitude regions, although this has not been a uniform process. For example, while the nordicity of Yellowknife decreased due to increasing government infrastructure and transportation access, the nordicity of a town like Schefferville increased because of a reduction in economic activity.

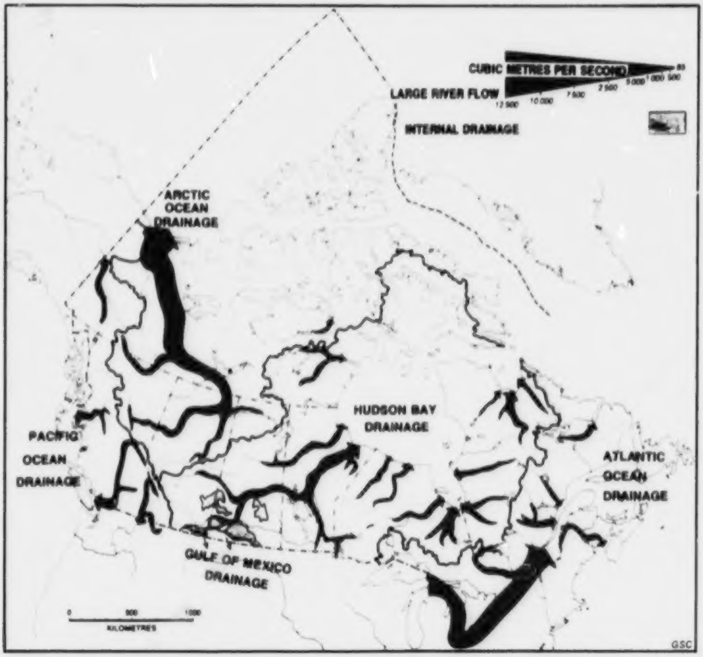

CUBIC METRES PER SECOND
LARGE RIVER FLOW
12 500
10 000
7 500
5 000
2 500
1 000
500
85
INTERNAL DRAINAGE
ARCTIC OCEAN DRAINAGE
HUDSON BAY DRAINAGE
PACIFIC OCEAN DRAINAGE
ATLANTIC OCEAN DRAINAGE
GULF OF MEXICO DRAINAGE
0
500
1000
KILOMETRES
GSC

Figure 1.2. Major drainage basins and annual river discharge *(after Greene, 1978).*

1.2.1.8. Guideline Definition

Thus the North could be defined geographically to include as little as the area north of the Arctic Circle or to encompass the entire land-mass of Canada and much of the northern United States. Because the geographical areas influenced by many of the physical processes discussed in this book lack spatial conformity, it was decided not to employ a stringent definition of northern hydrology but rather simply to use the political boundary of the 60th parallel to limit discussion where required (see Preface). Most of the physical descriptions contained in Chapter 1 use this arbitrary classification of the North. Importantly, however, some physical processes that dominate the hydrological environment of the northern latitudes have been researched most extensively in the more populated southern latitudes. Hence, in some cases, authors of other chapters have reviewed results from "southerly" studies to provide a more complete picture of research needs in the North.

1.3. NATURAL LANDSCAPE, CLIMATE AND HYDROLOGY

1.3.1. General Landscape

It is useful to consider the general scope and magnitude of the northern hydrological environment before discussing its specific hydroclimatic characteristics. The major physiographic regions north of 60°N include the Canadian Shield, the Interior Plains and the Western Cordillera (Figure 1.3).

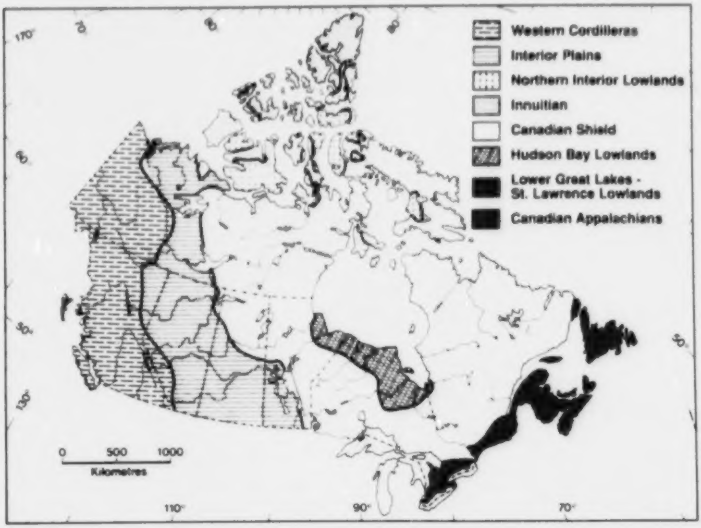
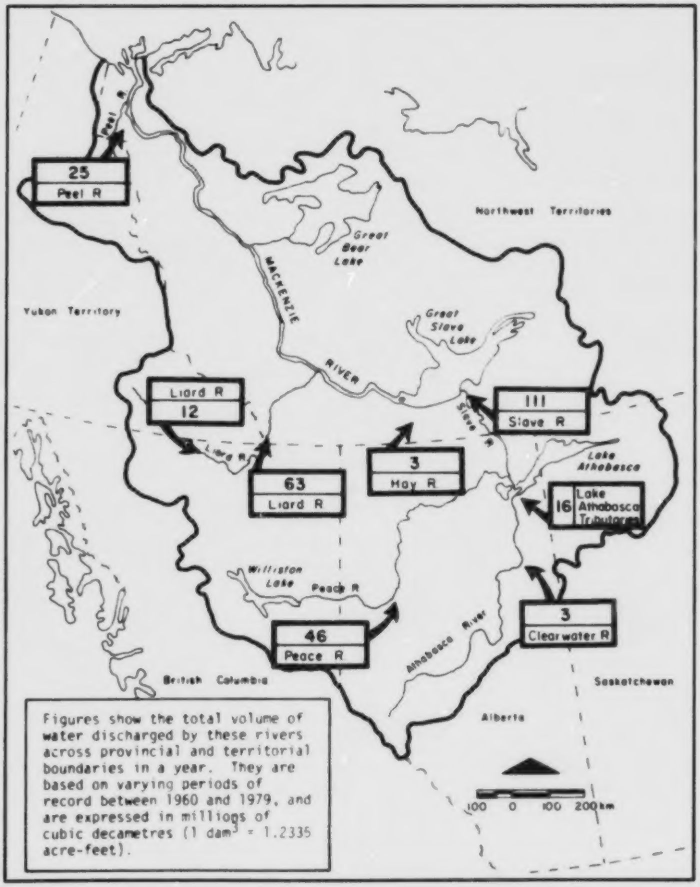

Western Cordilleras
Interior Plains
Northern Interior Lowlands
Innuitian
Canadian Shield
Hudson Bay Lowlands
Lower Great Lakes - St. Lawrence Lowlands
Canadian Appalachians
170°
0
500
1000
Kilometres
110°
90°
70°

Figure 1.3. Physiographic regions of Canada *(after Owen, 1967).*

All of the Shield was glaciated leaving a landscape comprised of ice-deepened valleys, eroded ridges, and extensive deposits of glacial till and lacustrine clays. It dominates the eastern and central portions of the Arctic Mainland where a maze of lakes and rivers blankets the surface because of the high proportion of exposed bedrock and generally poor drainage resulting from permafrost. True soil is generally absent and vegetation is largely of the tundra variety. Major river basins in this area include the Thelon (142,400 km^2), which flows into Hudson Bay, and the Back (106,500 km^2) that drains into the Arctic Ocean near Boothia Peninsula (Canada, Energy Mines and Resources, 1985). The Shield also extends into the eastern portions of the Arctic Archipelago, primarily Baffin Island.

The Arctic Archipelago comprises the islands north of the mainland and those in Hudson Bay. Seventeen of these islands are larger than Prince Edward Island, the smallest Canadian province, and five are an order of magnitude larger. The eastern islands are mountainous and covered with numerous glaciers and icefields. High plateaus dominate the central plains and lowland plains are most extensive in the western islands. Soils of the Arctic Islands are characterized by excessive stoniness and/or rock outcrops. Tundra vegetation dominates, comprised largely of lichen and dwarf species. Very limited vegetation occupies the most heavily-glaciated regions.

West of the mainland Shield country lie the Interior Lowlands, including the Mackenzie Lowlands to the north. They are largely comprised of a series of plateaus ranging in height from approximately 150 m in the east to 1200 m in the west (Owen, 1967). Glacio-lacustrine silt and clays dominate, especially in the eastern portions near Great Bear and Great Slave Lakes. These strata, in conjunction with permafrost, have led to the formation of extensive bog and muskeg wetlands. The west is bordered by the


25
Peel R
Peel
Northwest Territories
Great Bear Lake
Yukon Territory
MACKENZIE
RIVER
Great Slave Lake
Liard R
12
111
Slave R
Slave
Liard R
63
Liard R
3
Hay R
Lake Athabasca
16
Lake Athabasca Tributaries
Williston Lake
Peace R
46
Peace R
Athabasca River
3
Clearwater R
British Columbia
Saskatchewan
Alberta
100 0 100 200 km

Figures show the total volume of water discharged by these rivers across provincial and territorial boundaries in a year. They are based on varying periods of record between 1960 and 1979, and are expressed in millions of cubic decametres (1 dam^3 = 1.2335 acre-feet).

Figure 1.4. Transboundary flow of the Mackenzie River *(after James and Foerstel, 1983).*

higher alpine regions of the Western Cordillera from where flow Mackenzie River tributaries, such as the Arctic Red, Liard and Peel Rivers.

Immediately west of the Lowlands is the Mackenzie Mountains area, a portion of the Western Cordillera comprised of mountains, plateaus and plains. Rocks are largely of sedimentary origin. The other primarily alpine zone is that of the Coast Mountains, which is characterized by U-shaped valleys and extensive glacier systems, extending to the largest North American mainland icefield in the St. Elias Ranges. Bounded by the Mackenzie Mountain province in the north and east and the Coast Mountains to the south-west, is the Yukon Plateau. It consists of rolling upland dissected by valleys and some isolated mountains, the largest being the Pelly in the 2400 m a.s.l. range. The Yukon River dominates the central portions of the Plateau, draining approximately ¾ of the Yukon north-west into Alaska. The south-east portions of the plateau drain into the headwaters of the upper Liard River and ultimately into the Mackenzie River.

One other major plain area exists in the northern Yukon rimmed by a series of mountain chains. The major features include Eagle Plain and the Porcupine Plateau, a gently undulating plateau pock-marked by numerous lakes that drain to the west by the Porcupine River and ultimately the Yukon. By contrast, the remainder of the Yukon has very limited surface storage. The largely mountainous landscape means the Yukon has the smallest ratio of freshwater to total area (<0.01) of all the Canadian territories and provinces (Canada, Statistics Canada, 1987).

The dominant hydrological system in the North is the Mackenzie River. The catchment produces a flow comparable to that of the St. Lawrence and is the largest in Canada (1,805,200 km^2) including portions of both territories and three provinces (British Columbia, Alberta and Saskatchewan) (Figure 1.4). On a global scale, the Mackenzie is the 12th largest basin in size and the 14th largest river in flow (Encyclopedia Britannica, 1987). The mouth of the Mackenzie also contains Canada's largest active delta, approximately 200 km long and 65 km wide. Lake coverage is extensive ranging from 5% in the outer delta to 50% in the middle portions (Mackay, 1963).

1.3.2. Hydroclimatic Characteristics

1.3.2.1. Precipitation/Snowcover

Except for the mountainous parts of the Arctic Islands and West Coast regions, mean annual total precipitation north of 60°N is less than 400 mm (Figure 1.5). Rainfall inputs account for approximately 100-200 mm of total precipitation between 60°N and the arctic coast and above this point less than 100 mm (Figure 1.6). Except for the High Arctic and mountainous regions, where more than 50% of the total precipitation occurs as snowfall (Figure 1.7), precipitation is dominated by rainfall, primarily during the summer months. Precipitation data for the North, however, should be used very circumspectly because of the difficulties in obtaining accurate, representative measurements. Problems specific to the northern environment include a small measurement network, lack of alpine stations, particularly in areas of strong orographic activity, and pronounced gauge-catch deficiencies associated with snow and high winds.

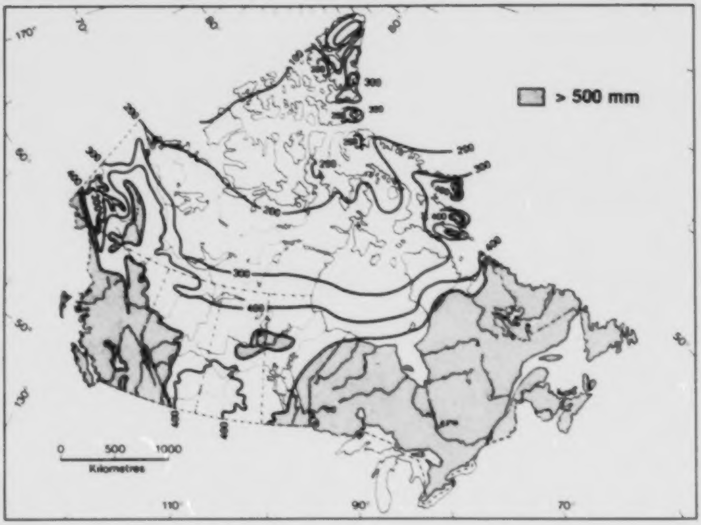
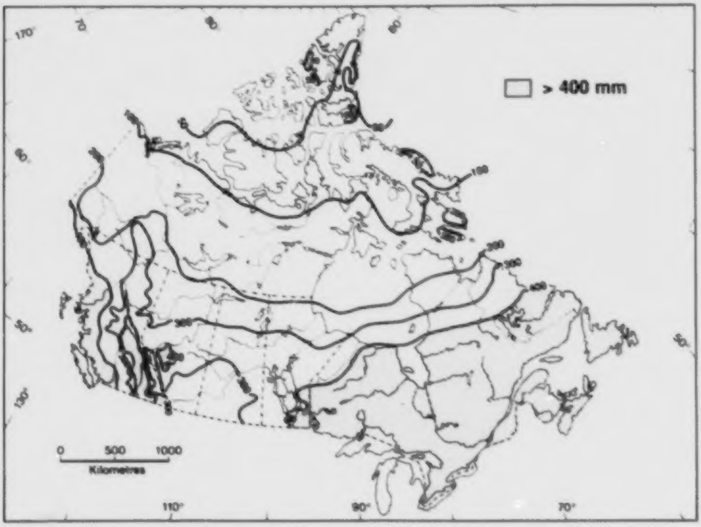

▨ > 500 mm
170°
70°
60°
50°
130°
110°
90°
70°
200
300
400
0 500 1000
Kilometres

Figure 1.5. Mean annual total precipitation (mm) 1951-1980 *(after Canada, Environment Canada, 1986a).*


☐ > 400 mm
170°
70°
60°
50°
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Kilometres

Figure 1.6. Mean annual total rainfall (mm) 1951-1980 *(after Canada, Environment Canada, 1986a).*

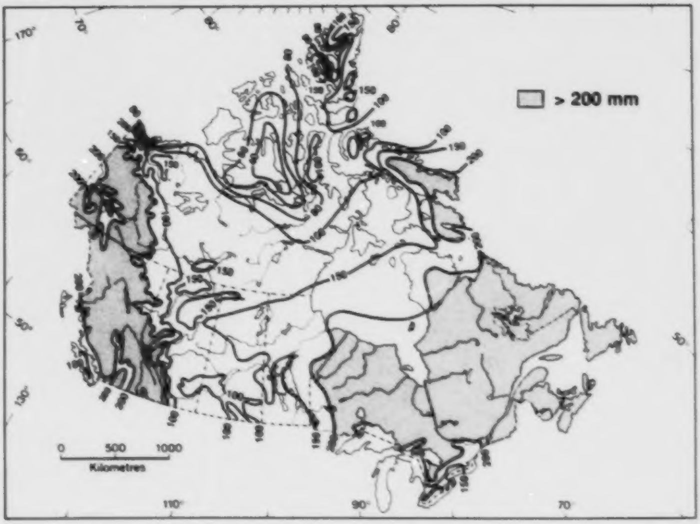


Figure 1.7. Mean annual total snowfall (cm) 1951-1980 *(after Canada, Environment Canada, 1986a).*

Much of the Arctic is essentially a polar desert because of the low precipitation, most of which is generated by infrequent frontal activity, usually at the leading edge of low-pressure systems. Convective activity is minimal because of the stabilizing effect of cold temperatures especially during the winter months when most local moisture supplies are ice-covered. Coastal areas experience a distinct precipitation increase during the open-water period. Orographic activity is responsible for the precipitation maxima in the Western Cordillera, such as in the St. Elias Mountains where annual total precipitation can exceed 1.5 m (over a metre as snow), and in the eastern, mountainous coasts of the Queen Elizabeth Islands and Baffin Island. In both alpine areas, the relatively large precipitation feeds extensive glacier systems.

Despite the small snowfall in the North, snowcover duration can be quite long. In general, all of the area north of 60°N becomes snow-covered by late-October (when ≥ 2.5 cm of snow occurs and remains more than 1 week) and can become snow-clear (when ≤ 2.5 cm of snow occurs and remains absent for 1 week or more) as early as the end of April (Findlay, 1978a) giving a North that remains snow-covered for at least 50% of the year (Figure 1.8). Actual depth of coverage, however, can be quite variable as evidenced by the mean maximum snow depths in Figure 1.9. While many high-alpine locations remain snow-covered all year and some of the Arctic Islands plains may be snow-free for as little as two months during the summer, areas of very low snowfall may have sporadic winter snowcovers never reaching more than 0.3 m.

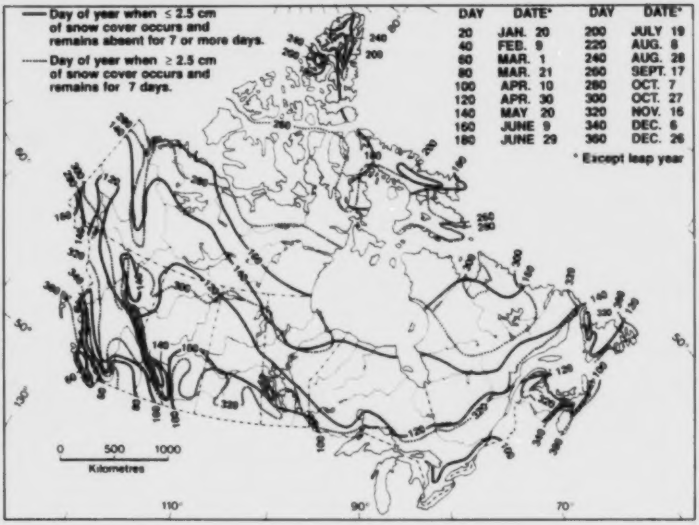

Day of year when ≤ 2.5 cm of snow cover occurs and remains absent for 7 or more days.
Day of year when ≥ 2.5 cm of snow cover occurs and remains for 7 days.
DAY DATE* DAY DATE*
20 JAN. 20 200 JULY 19
40 FEB. 9 220 AUG. 8
60 MAR. 1 240 AUG. 28
80 MAR. 21 260 SEPT. 17
100 APR. 10 280 OCT. 7
120 APR. 30 300 OCT. 27
140 MAY 20 320 NOV. 16
160 JUNE 9 340 DEC. 6
180 JUNE 29 360 DEC. 26
* Except leap year
0 500 1000
Kilometres
110°
90°
70°

Figure 1.8. Dates of formation and loss of snowcover *(after Findlay, 1978a).*

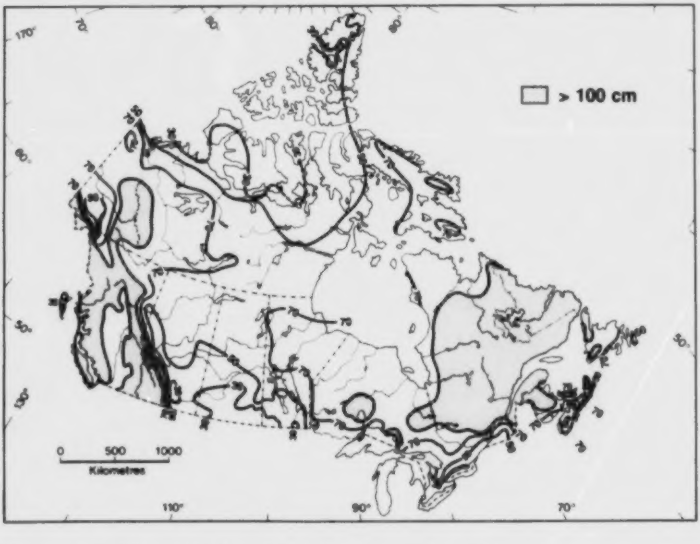

> 100 cm
0 500 1000
Kilometres
110°
90°
70°
70

Figure 1.9. Mean maximum depth of snow in cm *(after Findlay, 1978b).*

1.3.2.2. Drainage/Run-Off

Although northward-flowing rivers in Canada drain almost 85% of the total land mass (excluding drainage from the United States), the 1578 $km^3 \cdot yr^{-1}$ of water carried by these rivers is only 48% of the total annual run-off draining from Canada. Two reasons account for the large discrepancy. Firstly, the flow in some of Canada's more southerly rivers is greatly enhanced with run-off from an additional catchment area of approximately 577,000 km^2 contained within the U.S.A. The northward-flowing rivers receive an additional U.S. contribution of only ⅓ this area. Second, the southern rivers simply receive much more recharge from greater precipitation, particularly in the Great Lakes area and the Western Cordillera.

Much of the water in the northward-flowing rivers originates at latitudes below 60°N. The usable water supply over much of the North is actually quite small because of low precipitation and the storage effect of snow and ice, including permafrost. In fact, it is the permafrost layer that gives a false impression of abundant water in the North. Ponding of meltwater and the accumulation of precipitation produces extensive wetland zones which dominate the run-off regime for large areas of the North (Figure 1.10) (see Sections 1.3.2.3. and 1.4.1.4.).

Many of the catchments flowing into Hudson Bay yield more than 500 $mm \cdot yr^{-1}$ but significantly lower values are found for other areas north of 60°N. In general, run-off depths vary from approximately 50 to 250 mm, with the largest yields in some of the uplands, eastern Arctic Islands and

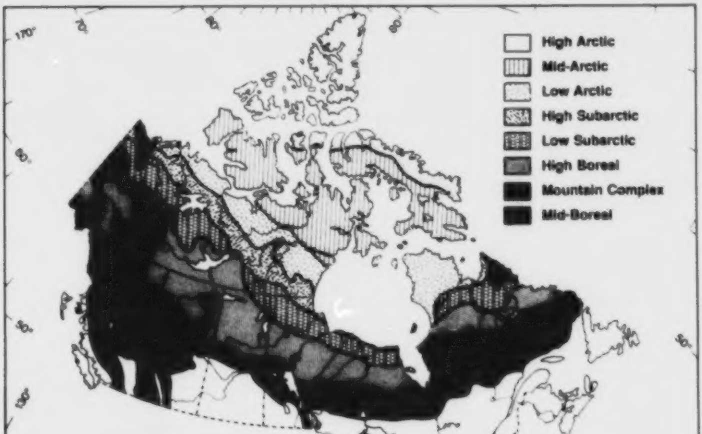


Figure 1.10. Wetland regions of northern Canada *(after Jackson, 1989).*

much of the Yukon (Laycock, 1968). The yield of the entire Mackenzie River catchment is only 175 mm·yr^{-1} (after Canada, Energy Mines and Resources, 1985) despite extensive contributions from south of 60°N and the western Cordillera.

Within the mountainous western regions and more southerly areas, rainfall run-off is responsible for the annual hydrographic peaks. In areas of low relief, such as muskeg and peatlands, basin response is slowed and the peaks become flattened. Snowmelt becomes increasingly important with latitude until ultimately almost all of the annual streamflow within the High Arctic results from snowmelt. Infrequent and usually small-magnitude rainfall events are insufficient to generate run-off in these areas except immediately after snowmelt when the ground is saturated (e.g., Woo, 1983). Total run-off in high northern basins is often more than first expected from the low precipitation. This is because evaporative losses are low and infiltration is negligible due to the effect of permafrost.

Low flow on northern rivers occurs during the winter when ground-water recharge is minimal and significant quantities of water are being converted to ice (growth within the channel system) and placed in storage (through backwater effects; see Section 5.3.2.). Flows generally decrease with latitude and permafrost extent; minimum mean monthly discharge rates ranging from about 0.002 to approximately 0.005 m^3·s^{-1}·km^{-2} in the more-southerly discontinuous zone to near zero in the continuous zone (Figure 1.11). Although zero surface discharge does occur on some streams, some flow may still occur through unfrozen gravel beds. Flow values greater

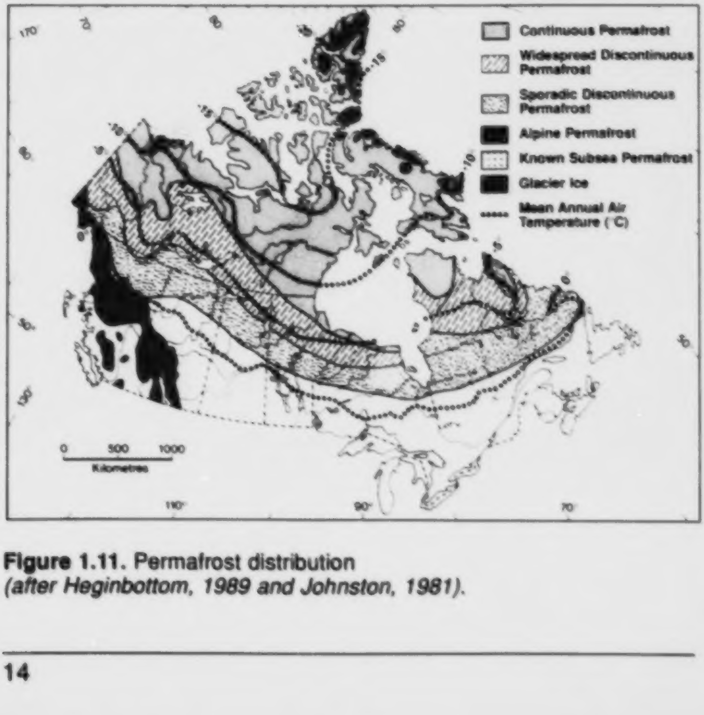

Continuous Permafrost
Widespread Discontinuous Permafrost
Sporadic Discontinuous Permafrost
Alpine Permafrost
Known Subsea Permafrost
Glacier Ice
Mean Annual Air Temperature (°C)
0 500 1000
Kilometres
170°
110°
90°
70°

Figure 1.11. Permafrost distribution *(after Heginbottom, 1989 and Johnston, 1981).*

than 0.005 $m^3 \cdot s^{-1} \cdot km^{-2}$ usually reflect significant surface storage within the catchment (van Everdingen, 1987).

1.3.2.3. Surface Storage

Over 750,000 km^2 of Canada's land mass is covered by freshwater, eighteen percent of which is found north of 60°N, 133,300 km^2 within the Northwest Territories and 4,480 km^2 within the Yukon (Canada, Statistics Canada, 1987). Most is concentrated in a wide band between Great Bear and Great Slave Lakes and in the middle portions of the District of Keewatin, immediately west of Hudson Bay (Canada, Energy Mines and Resources, 1973; Plate 7-8). Twenty-eight large lakes in excess of 600 km^2 or almost 40% of the Canadian total, lie within the N.W.T. (Canada, Statistics Canada, 1987). Notably, two of these lakes, Great Slave and Great Bear, are each larger than all but two of the Canadian Great Lakes.

These freshwater areas do not include the extensive glacierized areas that total in excess of 150,000 km^2 on the Arctic Islands and an additional 15,000 km^2 on the territorial mainland; an area that exceeds 75% of the Canadian total (Ommanney, 1970).

Much of the territories is also covered with wetlands (including peatlands and adjacent shallow water < 2 m in depth) (Figure 1.10). Although small areas occur on the west coast of Hudson Bay and in the eastern Arctic Islands (Banks and Victoria Islands), the greatest concentration occurs along the Mackenzie River valley. The total wetland area in the North is 293,040 km^2; exceeding that contained in any of the southern provinces and accounting for 23% of Canada's total wetland area (Canada, Energy Mines and Resources, 1986).

1.3.2.4. Permafrost

Permafrost, ground that is continuously below 0°C for two years or more (Canada, National Research Council, 1988), covers approximately one-quarter of the land surface of the world and one-half of Canada (U.S. National Research Council, 1983). Most present permafrost originated during the Pleistocene (Washburn, 1980) with the thickest developments found in unglaciated areas. Permafrost is of post-glacial origin in most formerly glaciated areas. For much of northern Canada, it plays a significant role in the terrestrial portion of the hydrological cycle because it restricts moisture exchanges between surface water and deep ground water.

Figure 1.11 shows the approximate distribution of permafrost within Canada. For the most part, permafrost isolines are broadly extrapolated from a meagre spatial data set. Although they suggest that sharp boundaries exist, the thickness, temperature and horizontal extent of permafrost undergoes a gradual transition northward. The boundary between continuous and discontinuous permafrost approximates the position of the -8.5°C mean annual air isotherm. The southern limit of permafrost is generally accepted as coinciding with the -1°C mean annual air isotherm, although isolated patches exist further south, particularly in peatlands. In the continuous zone, permafrost thickness generally increases with latitude from approximately 100 m at the southern limit to 1000 m in the far North. Continuous permafrost is also found in the Western Cordillera, even occupying valley bottoms at the northern limit (Figure 1.11), but increasing in elevation with decreasing latitude to approximately 2000 m a.s.l. at 49°N (Brown, 1978).

Permafrost distribution is controlled by a combination of climatic, geological, hydrological, topographical and botanical factors. The anomalous southern zones of permafrost primarily exist because of the insulating effect of overlying organic material. Its low thermal conductivity, seasonal variation in conductivity, and tendency to promote an evaporation regime which promotes cool summer temperatures, all serve to create an environment where permafrost can exist even where the mean annual air temperature is above freezing.

Of all the factors affecting permafrost, water is the most significant. In liquid form, the combination of high absorption of solar radiation, ease of turbulent mixing and high thermal capacity serve to develop significant summer heat storage and hence, significant ground warming. During the winter, most large lakes and rivers with appreciable flow do not freeze to the bed and hence, many rivers and lakes have unfrozen areas or taliks beneath them which permit the linking of the surface- and ground-water flow systems. In the solid form, as snowcover, water has dual conflicting roles in affecting permafrost. Because of its low thermal conductivity, snow inhibits ground freezing during the cold winter months but its high albedo delays thawing and warming of the underlying ground in the spring.

In contrast to the effect of large water bodies on permafrost occurrence, the restriction of infiltration and recharge of water by permafrost over areas of low relief can lead to the formation of extensive muskegs and ponds. Even large lakes, termed thermokarst or thaw lakes, can owe their existence to land subsidence brought about by permafrost degradation. In many cases, the location of surface drainage-divides and the presence of perched water-tables depend on permafrost topography. Catastrophic drainage of some lakes has been linked to the melting out of such permafrost divides. Long-term contributions to run-off from changes in the mass of ice stored in permafrost, however, is virtually unknown.

Overall basin response in permafrost regions is controlled primarily by the depth of a surface active layer which may exist for as little as 2 months in the far North. With a fully-developed active layer, recession times can be much greater than in more temperate regions (Church, 1974). In general, however, the unsaturated zone within the active layer is relatively thin and offers only limited storage. In sloped terrain underlain by permafrost, interflow and run-off potential is much higher than in similar non-permafrost environments.

1.3.2.5. Ground Water

According to Hess (1986; Table 4.2), total 1981 water use in the territories was 34.9 x 10^6 m^3 of which 37% was supplied from ground-water sources. Within the Northwest Territories, where almost half of the total water supply comes from ground water, it was almost entirely (99.7%) used by the mining industry (primarily in mine dewatering). By contrast, virtually all of the ground water extracted in the Yukon (22% of total water supply) is consumed by municipal and rural water users. Well over half of all Yukon residents rely on ground-water supplies whereas in the N.W.T. this figure is less than 1%.

To a large degree, the supply, availability, distribution and quality of ground water in the North is controlled by the presence of permafrost. This impermeable layer controls both the recharge to underlying aquifers and the discharge to surface-water systems. The resulting ground-water systems hold either (a) suprapermafrost water, (b) intrapermafrost water or (c) subpermafrost water. Suprapermafrost water is the most widely spread, occupying the seasonally-formed active layer and unfrozen zones beneath lakes and rivers. Water obtained from such other sources often offer only a seasonal supply, are easily contaminated and, in the case of active-layer water, are high in organic content. Intrapermafrost water exists within thawed zones between perennially-frozen layers. Common locations include alluvial layers near rivers, abandoned river channels and glacio-fluvial material (Owen, 1967). Subpermafrost water is found beneath permanently frozen ground, usually in either bedrock or deep glacio-fluvial deposits.

Alluvial and glacio-fluvial deposits, which can contain any of the three permafrost-water types, provide good ground-water supplies for many northern communities within the discontinuous permafrost zone (Brandon, 1965). Progressing northward, such deposits become less accessible with increasing permafrost, although deep bedrock aquifers can produce high-quality water, especially within the high-recharge zone adjacent to the western mountains. For the Canadian Shield region, ground-water flow within the crystalline rock is largely restricted to weathered zones, fracture systems and faults, and it is the alluvial and glacio-fluvial deposits that may offer the only source for ground water over large areas of the Shield (van Everdingen, 1974). On the Arctic Islands, the low precipitation and enhanced concentration of permafrost further limit the availability of ground water.

1.3.2.6. Evaporation/Evapotranspiration

Of the major hydrological components, evaporative processes are the least understood, especially in northern areas. Figure 1.12 shows the mean annual lake evaporation (den Hartog and Ferguson, 1978a) decreasing from approximately 250-400 mm at 60°N to less than 100 mm in northeastern Keewatin and the central portions of the Arctic Islands. Notably, these figures are based on Class A evaporation pan data supplemented by calculations from additional climatological data. Given that the pan network in the entire N.W.T. is comprised of only five stations (Latham, 1988), extreme care must be used in any interpretation of the data. Comparison of Figures 1.5 and 1.12 suggests that for much of the North, lake evaporation is equal to or greater than 50% of precipitation. Based on the available pan data for the N.W.T., Latham (1988) showed that lake evaporation is even greater than precipitation.

Such comparisons, however, neglect the role of transpiration and evaporation from other surfaces. Figure 1.13 illustrates values for evapotranspiration calculated by den Hartog and Ferguson (1978b) from the data in Figure 1.12 by first assuming potential evapotranspiration equal to lake evaporation and then assigning a ratio between potential and actual evapotranspiration for different surface types. Optimization of these values with those for

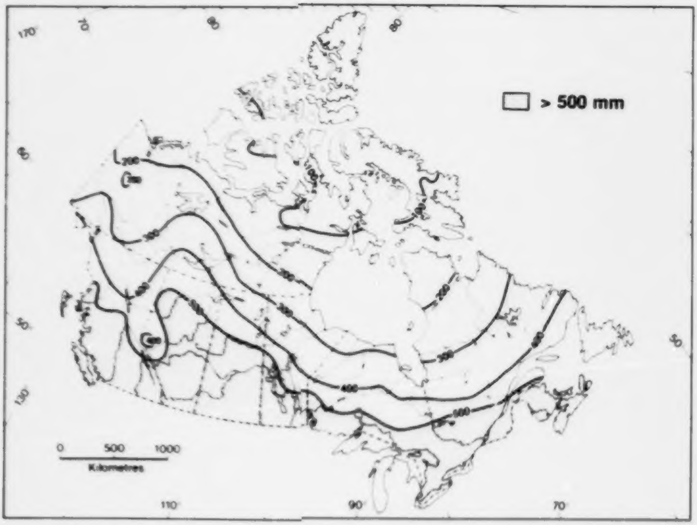
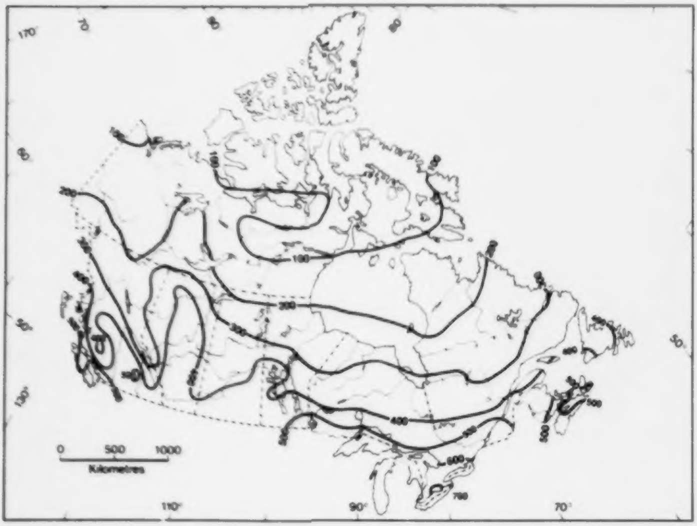

170°
> 500 mm
200
300
400
500
0 500 1000
Kilometres
110°
90°
70°

Figure 1.12. Mean annual lake evaporation *(after den Hartog and Ferguson, 1978a).*


170°
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500
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0 500 1000
Kilometres
110°
90°
70°

Figure 1.13. Derived evapotranspiration *(after den Hartog and Ferguson, 1978b).*

run-off and precipitation lead to the final values shown in Figure 1.13. Again, the major weakness of this data for the North is simply the inadequate measurement network. Chapter 8 presents more detailed evaporation/evapotranspiration data and discusses the importance of evapotranspiration relative to that of precipitation for a variety of climatic conditions.

In general, evaporation is greatest during the summer, especially in the areas of low relief characterized by numerous bogs and lakes. In such shallow-water systems, the evaporation rate is closely controlled by variations in daily energy input. On larger, deeper water bodies, heat storage becomes important to evaporation. Thus in the fall months, it is only the deeper lakes with appreciable heat stored from the summer months that continue to produce significant evaporation.

A major complication in the extensive wetland areas is the presence of peat which controls the availability of moisture for the evaporation process. It is also evaporation from saturated peat that protects the underlying permafrost from warming by consuming the heat from incoming short-wave radiation (Williams, G.P., 1974). Moving northward, and out of the wetland and forested zones, transpiration losses generally decrease because of the lower vegetation density and the increasing proportion of mosses and lichens which do not transpire. However, the relative importance of evapotranspiration in the overall water balance tends to increase because of the more rapid northward decrease in precipitation.

Moisture loss to the atmosphere can also occur through snow sublimation. Although little information is available, it is likely to be most significant during the spring high-radiation period, especially in high-wind, low-relief areas and, on leeward slopes of the Western Cordillera and in the mountainous zones of the Arctic Islands affected by föhn-like winds. Because heat available to drive the evaporation process is usually small and available for only short periods in the far North, the resulting evaporation should also be small, but perhaps large relative to the total snowfall.

1.3.2.7. Freshwater Ice Cover

Freshwater ice seasonally covers most lake and river systems in Canada, ranging from periodic skims of ice in the more southerly temperate regions to mean thicknesses in excess of 2 m on high-latitude northern lakes and rivers (Figures 1.14 and 1.15). Sufficient ice develops that waterways are unnavigable for over 100 days within the southern portions of the provinces and for over 180 days within the northern portions of the provinces and the territories (Bates and Bilello, 1966). The influence of latitude and maritime effects are apparent in Figures 1.16 to 1.19 which depict mean dates of freeze-over and completion of ice clearance on lakes and rivers.

In general, the size of the water body determines the rate of freeze-up, with small lakes requiring less than 2 weeks from first ice appearance to complete freeze-over, and large lakes, such as Great Bear and Great Slave, as long as two months (Allen, 1977). Freeze-up on rivers differs from that of lakes (Chapter 5) and occurs later within the same region. Duration of the river-ice season is also less than that of lake ice because river ice is first to break-up in the spring. The northward advance of river-ice break-up in the territories occurs on average at a rate of approximately 0.3 degrees of

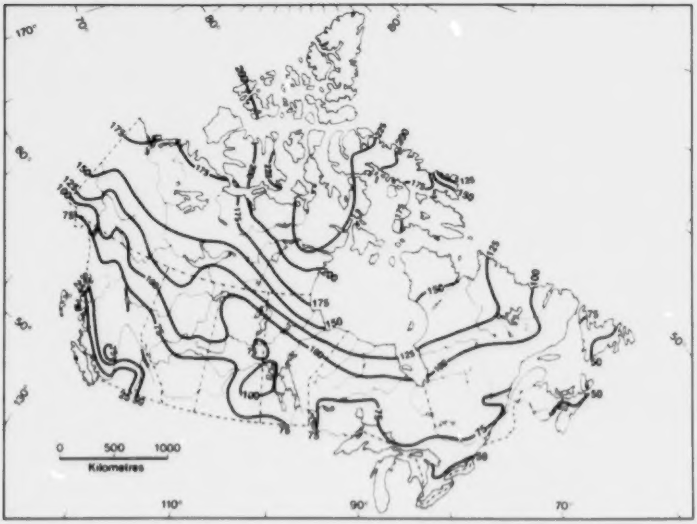
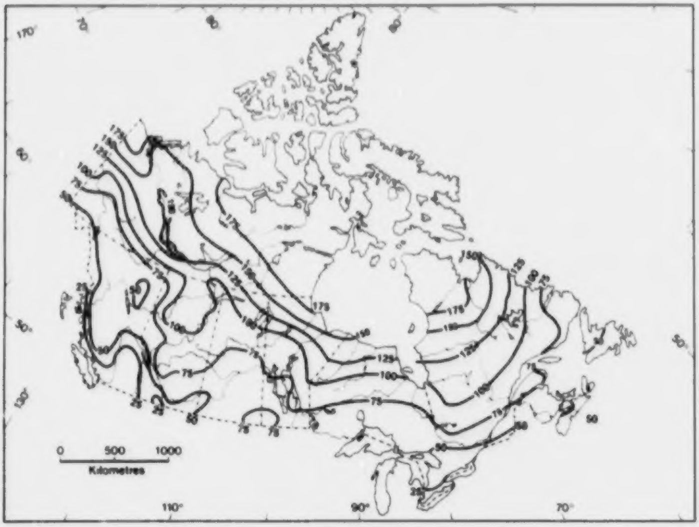

170°
70°
80°
80°
60°
50°
130°
50°
0 500 1000
Kilometres
110°
90°
70°
25
50
75
100
125
150
175
200

Figure 1.14. Mean maximum ice thickness (cm) in lakes *(after Allen, 1977).*


170°
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130°
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0 500 1000
Kilometres
110°
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25
50
75
100
125
150
175

Figure 1.15. Mean maximum ice thickness (cm) in rivers *(after Allen, 1977).*

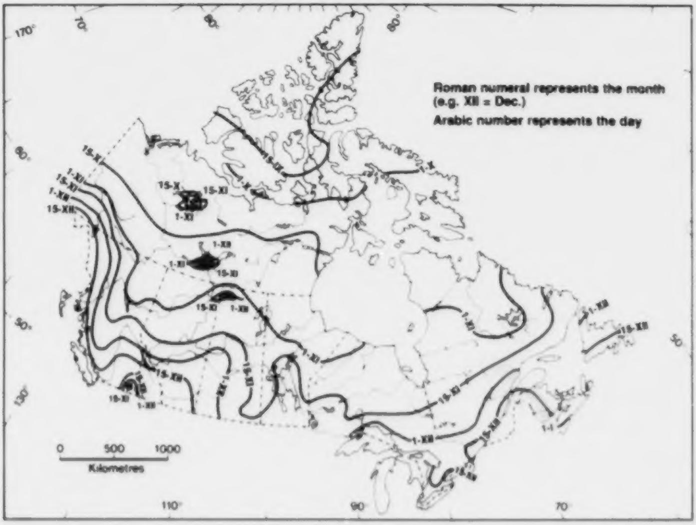
Roman numeral represents the month
(e.g. XII = Dec.)
Arabic number represents the day
170°
70°
80°
80°
60°
50°
130°
50°
0 500 1000
Kilometres
110°
90°
70°
15-XI
1-XII
15-XI
1-XII
15-XII
15-X
15-XI
1-XI
1-XII
1-XI
15-XI
15-XI
1-XII
1-XII
15-XII
15-XII
1-I

Figure 1.16. Mean freeze-over date on lakes *(after Allen, 1977)*.

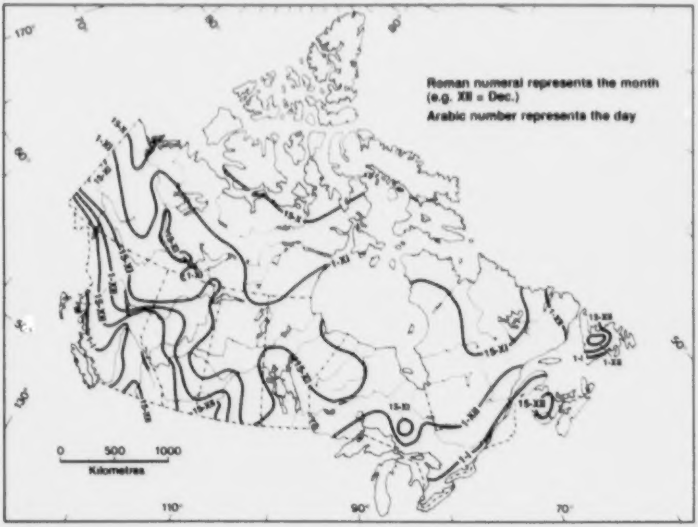
Roman numeral represents the month
(e.g. XII = Dec.)
Arabic number represents the day
170°
70°
80°
80°
60°
50°
130°
50°
0 500 1000
Kilometres
110°
90°
70°
15-X
1-XI
15-XI
1-XII
15-XII
1-XI
15-XI
15-XII
1-XII
15-XI
1-I
1-XII
15-XII
1-I

Figure 1.17. Mean freeze-over date on rivers *(after Allen, 1977)*.

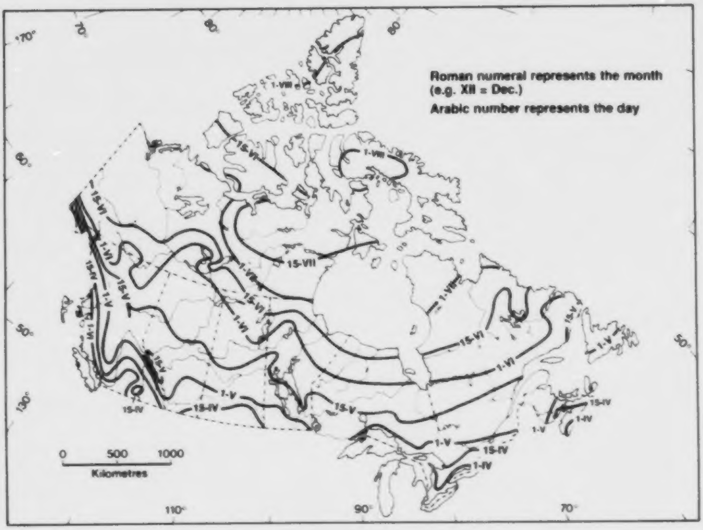
Roman numeral represents the month
(e.g. XII = Dec.)
Arabic number represents the day
1-VIII
15-VII
1-VII
15-VI
1-VI
15-V
1-V
15-IV
1-IV
0 500 1000
Kilometres
170°
70°
80°
60°
50°
130°
110°
90°
70°

Figure 1.18. Mean ice-free date on lakes *(after Allen, 1977).*

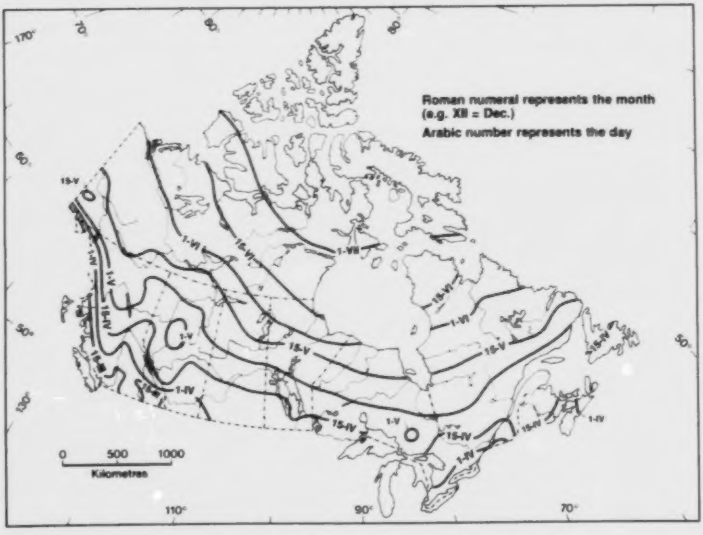
Roman numeral represents the month
(e.g. XII = Dec.)
Arabic number represents the day
1-VII
15-VI
1-VI
15-V
1-V
15-IV
1-IV
0 500 1000
Kilometres
170°
70°
80°
60°
50°
130°
110°
90°
70°

Figure 1.19. Mean ice-free date on rivers *(after Allen, 1977).*

latitude per day (Prowse and Onclin, 1987). Complete clearance of lake ice does not always occur in the far North and multi-year ice periodically develops on some lakes because of the brevity of melting seasons. Similarly, some forms of ice cover can remain on streams and rivers until late summer and, in some years, form multi-year accretions. Such covers are found where ground-water flow has contributed to the development of exceedingly thick surface-ice accretions, termed icings, as described in Section 4.3.4.3.

1.3.2.8. Data-Collection Networks

Although the above descriptions and figures on the northern hydrological environment appear detailed, much of the information is based on extrapolation from meagre data bases, both in terms of length-of-record and spatial coverage. As will be described in subsequent chapters, much of the detailed knowledge of northern hydrological processes has been gained from site-specific research studies. Many broad generalizations about northern hydrological regimes, however, have been based on data collected from the standard hydrometric- and climatic-station network. Because of the importance of these networks to both large-scale research studies and to water planning and management, they are briefly reviewed here.

As of 1989, there were 181 hydrometric stations (note: all discussion below refers to hydrometric stations providing discharge records) (Canada, Environment Canada, 1990; data by personal communication from IWD staff, Yellowknife and Whitehorse) and 115 climate stations operating within the combined Yukon and Northwest Territories. Representing 86% of the total Canadian area above the 60th parallel, the N.W.T. contains most of the hydrometric and climate stations (64% and 83% respectively). Because of the vastness and remoteness of the N.W.T., it has been the focus of a number of network reviews. For example, in a review of climate stations in the N.W.T., Acres (1982) concluded that the distribution and density of such stations was particularly poor and that additional remote stations were needed especially in the Keewatin between Great Bear and Baker Lakes, and in the interior of the Arctic Islands. In the case of hydrometric stations, Acres (1982) noted that the overall hydrometric network was simply inadequate for the derivation of a good hydrological data base and that the distribution of the network was somewhat uneven, partly due to a concentration of stations along the Mackenzie River system.

Figures 1.20 and 1.21 show the distribution by basin area and length-of-record of the hydrometric stations north of 60°N. According to the World Meteorological Organization guidelines (WMO, 1981), the recommended hydrometric-station density for polar zones ranges from one per 5,000 to 20,000 km^2, equally divided among catchments greater than and less than 10,000 km^2. Within the territories, the 1989 overall ratio is approximately 1:22,000 km^2 with approximately 61% (of stations with defined drainage basin area) located on the smaller catchments. Importantly, however, the ratio is 1:8,000 km^2 for the Yukon and 1:29,000 km^2 for the N.W.T. with 68% and 58% (respectively) of the stations located on catchments less than 10,000 km^2. As suggested by Wedel (1986), it could be argued that a lower station density is sufficient for the larger catchments in the N.W.T. since many are predominantly permafrost with low precipitation and

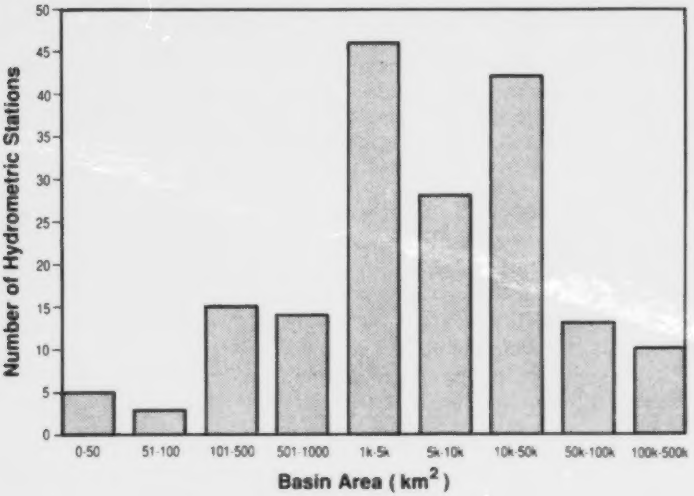

Figure 1.20. Number of hydrometric stations in the North by basin area.

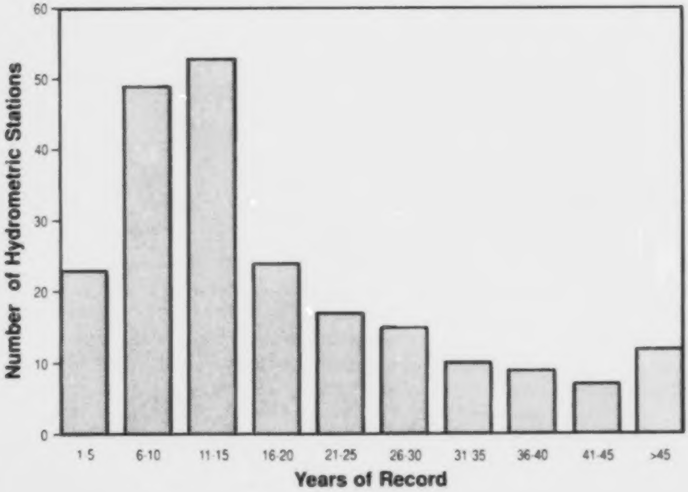

Figure 1.21. Length of record of northern hydrometric stations.

evapotranspiration. There is no doubt, however, that if development of the territories continues, the baseline data network will have to increase to ensure proper water management. Unfortunately, even the current record length of many stations is insufficient for probabilistic analysis of various flow regimes. For example, only 38 (equally divided between the Yukon and N.W.T.) or 21% of the discharge stations have records longer than 30 years in length. Even these long time-series may have serious omissions, inconsistencies and inadequacies, particularly during the freeze-up and break-up periods (e.g. Ozga, 1988). Typical ice problems affecting flow measurement are reviewed in Sections 5.3.4. and 5.4.4.

As part of the Water Survey of Canada hydrometric network, periodic ice-thickness measurements are made at most sites, but the real network of ice-reporting stations that document detailed ice information, such as times of freeze-up/break-up and maximum ice thickness, is much smaller. As of 1984, the network was comprised of only 29 river and 61 lake locations (Canada, Environment Canada Service, Atmospheric Environment Service, unpublished data summary). The snow-survey network in northern Canada is also meagre. The most recent summary report (Canada, Environment Canada, 1986c) lists only 74 official reporting snow surveys; 20 within the Northwest Territories and 54 in the Yukon.

There is currently no co-ordinated glacier-survey program operating in northern Canada, although some long-term studies have been carried out by government departments (Geological Survey of Canada, Energy Mines and Resources; National Hydrology Research Institute, Environment Canada) and some universities (University of British Columbia, University of Ottawa, Trent University, University of Colorado, etc.). More details are provided in Chapter 6.

Water-quality monitoring in the North, is conducted both at standard hydrometric stations by the Water Quality Branch (WQB) of Environment Canada and at approximately 70 industrial and municipal water-use operations licensed by the Department of Indian Affairs and Northern Development (DIAND). The WQB network includes only 20 stations and has a limited historical data base, particularly during the spring break-up period. Furthermore, the range of parameters is small and not oriented toward detecting environmental contaminants. Further details about both networks are provided in Sections 7.2.2 and 10.6.2.

1.4. HYDROLOGICAL CLASSIFICATIONS

A number of attempts have been made to classify the North into distinct hydrological regimes using information gathered from the northern data-collection networks and from the myriad of case studies.

1.4.1. Church (1974) Classification

One of the most commonly used classifications is that of Church (1974) who identified four distinct regions based on the source and timing of run-off as determined from annual hydrographs. These included: arctic-nival, subarctic-nival, proglacial and muskeg.

1.4.1.1. Arctic-Nival

The arctic-nival region occupies the continuous-permafrost regions where deep infiltration is impeded by perennially-frozen strata and hence, base flow and winter flow to rivers is low. Spring snowmelt forms the major flow event of the year. Summer peaks can also result from rainfall events, but since precipitation is generally low, flood-level flows are usually generated only on very small basins.

1.4.1.2. Subarctic-Nival

This regime is also frequently dominated by spring snowmelt events, although spring rainstorms can produce larger discharges, especially for rivers with mountainous headwaters such as the main rivers flowing eastward from the Western Cordillera. Peak water levels are, however, often due to backwater effects of ice jams that accompany spring run-off. Groundwater flow contributions are larger than the arctic-nival because of lack of continuous permafrost. Complete freezing of the watershed does, however, preclude winter flow on some far northern rivers. For rivers with headwaters in the southern discontinuous-permafrost zone, flow may continue throughout the winter with much of the base flow derived from unfrozen channel gravels or, in the Cordillera, from springs (see Section 3.2.1. for additional "spring" regime classification).

1.4.1.3. Proglacial

Within the proglacial hydrological regime, snowmelt produces a spring rise in discharge but flow continues to rise throughout the summer as progressively higher zones of the glacial system contribute run-off. A similar hydrographic response can be created by the melt of late-lying or semi-permanent snow banks at higher altitudes (Marsh and Woo, 1981). Basins dominated by melt produced from such snow banks exhibit a proglacial-like summer peak rather than the spring peak associated with arctic-nival regimes.

Ponding of meltwater in glacier-fed systems can lead to the formation of ice-dammed lakes and ultimately to an exceptionally large glacier-burst flood or jøkulhlaup when such impoundments break. Within some proglacial basins, the combination of spring snowmelt, summer glacier/snowmelt and rainfall can produce highly complex hydrograph behaviour from which it is difficult to separate the various sources temporally.

1.4.1.4. Muskeg

The muskeg regime applies to large areas of low relief characterized by poor drainage. Church (1974) suggests that run-off attenuation is high because of the large water-retaining capacity of muskeg and its large flow resistance. Similar effects occur in tundra zones where thick mats of grasses or shrubs and hummocky surfaces restrict flow and in lowland tundra where numerous small water bodies attenuate flow. The variability of annual peak flows may be reduced by the presence of small lakes and extensive wetlands (Woo, 1986).

1.4.2. Network Classification

Separate classifications of hydrological regions have been conducted for the Yukon and N.W.T. by government agencies in each of the territories. Within the Yukon, four zones have been identified: Western Mountains, Interior, Northern and Arctic (Janowicz, personal communication, 1990; Figure 9.1). In general, hydrological response differs between the north and south Yukon because of permafrost effects which increase with latitude (Janowicz, 1986). Because the density of hydrometric stations decreases significantly above approximately 65°N (Ogilive Mountains), most is known about basin response in the southern portions of the Territory. Streams of the Western Mountain region are characterized generally by a rapid increase in flow during the early summer from low-elevation snowmelt followed by a late-summer peak from glacial melt. This region, however, can be further subdivided into streams draining the St. Elias and Coast Mountains; the former being more extensively glacierized and exhibiting a flashy run-off response. The smaller glacierized area of the Coast Mountains, in combination with longer stream lengths and extensive lake systems produces a smoother, less variable, run-off response. Discharge on other, southern Yukon streams tends to: rise rapidly during the spring due to snowmelt, peak usually in June, and produce several weeks of sustained high flow during the summer from rainfall (Janowicz, 1986).

Since Chapter 9 contains a detailed review of hydrological classifications developed for the N.W.T. only a brief review is presented here. As a starting point for assessing the northern hydrometric network in the N.W.T., two general classifications were devised. The first, defined four basic regions based on a set of hydrological parameters including: major drainage divides, relief, geology, soil and vegetation cover, amount of lake storage and climatic variability (Acres, 1976). These included: the Cordilleran Region, the Interior Plains Regions, the Canadian Shield Region and the Arctic Islands (Table 9.1). A subsequent report (Acres, 1982) noted that improvements in the data base over the six-year intervening period did not justify further subdivision. In 1983, however, another classification based on the same criteria defined a total of six regions, achieved by subdivision of the Canadian Shield Region into three (Table 9.2; Canada, Environment Canada and Indian Affairs and Northern Development, 1983).

Perhaps the most detailed and comprehensive classification of northern hydrological regimes was proposed by Wedel (1986), who classified the N.W.T. (and adjacent similar areas) into 60 hydrological regions based on a broad set of physiographic and vegetational units (Section 9.2.4.). Vegetation was used as an integrator of the variability in precipitation, soil permeability and evapotranspiration. Importantly, only six regions were considered to have sufficient hydrometric stations to permit the satisfactory development of models for use in flood, annual yield and low-flow analysis. Moreover, Wedel (1986) noted that, even if the then rather ambitious N.W.T. network expansion was to be completed, there would remain 16 hydrological regions in the N.W.T. with no hydrometric program whatsoever. Wedel (1986) also stressed the need to direct any future data-collection efforts towards regions with existing or potential development. Areas of future concern for water resources, including that presented by development, are reviewed next.

1.5. DEVELOPMENT AND WATER USE

Water is perhaps the single most important resource in northern Canada. It holds significant potential for future developments activities and is linked inextricably with other resource-related developments. Before reviewing the present and future developments that might impact on water resources, it is useful to consider the forces that drove the historical development of the North.

1.5.1. Historical Resource Development

Canada's entire history of northern exploration has been driven largely by an expectation of wealth from the northern frontier. As stated by V. Stefansson, the famous northern explorer and scientist: "There is no northern boundary beyond which productive enterprise cannot go till North meets North on the opposite shores of the Arctic Ocean as East has met West on the Pacific." (Stefansson, 1924, p.19). Pushing back the northern frontier has also been a periodic rallying-cry for political campaigns; the most notable being the promises made by John Diefenbaker of "... a new Canada -- A Canada of the North" in which project assistance would be made available for "... the development of the national resources for the opening of Canada's Northland" (Diefenbaker, 1958, p.6-7). He likened the opportunity to settlement of the west and believed it would give Canadians a new sense of unity. This appears to have been the case since he subsequently captured a huge electoral victory from the largely, southern electorate.

The spirit and intensity of early northern development was due in part to a philosophy of climatic determinism in which the path of world power was considered to be continually moving northward into cold more rigorous northern regions. As late as 1958, one of the official handbooks of the federal government reiterated this philosophy: "It is a curious fact that civilization has been expanding northward ever since the dawn of history... For thousands of years, society has been converging from both sides of the world toward a common centre. That centre is the Arctic." (Canada, Northern Affairs and National Resources, 1958, p.35). Much of the early optimism for northern development focused on the use of the northern river systems, particularly that of the Mackenzie River. Such optimism is reflected in a 1988, Canadian Senate Committee report which stated that there were "650,000 square miles of northland suitable for potatoes, 407,000 for barley, and 316,000 for wheat. The potential for agriculture was said to stretch all the way to the Arctic Ocean along the Mackenzie Valley..." (Page, 1986, p.10).

Much of our scientific knowledge of the North is a by-product of commercial development and resource exploitation. Historically, only three major northern enterprises have produced important net profits on a national or international scale: the fur trade from 1670 to about 1920, whaling in the 1800s, and mineral development in the Yukon 1898-1910. More recently, mineral and petroleum development have been the major economic contributors to the Yukon and N.W.T., but the net income generated in the regions has remained less than inputs to the system primarily through: government services, subsidies, welfare and military expenditures (Adams *et al.*, 1987). From 1968 to 1985, large private and public investments were made in the North for the exploitation of petroleum resources. This activity assisted and necessitated an improved understanding of the northern environment, especially water resources. One example was the compendium of reports on the hydrological aspects of northern pipeline development produced for the Environmental-Social Program, Northern Pipelines of the Task force on Northern Oil Development, Government of Canada; a data base that formed part of the baseline information for the subsequent Mackenzie Valley Pipeline Inquiry conducted by Justice Thomas Berger (Berger, 1977).

During the 1980s, there was a continued recognition that, other than isolated resource-extraction industries, development would concentrate along the major river systems, particularly the Mackenzie River (Figure 1.22; e.g., James and Foerstel, 1983). Recognition of this potential led to the production of two major water-resource management studies for the Mackenzie (Mackenzie River Basin Committee, 1981a; see Table 10.1 and 10.2.) and Yukon River basins (Yukon River Basin Committee, 1984). These reports have set the stage for most water-management strategies for the North and many of the data and research requirements.

1.5.2. Present Resources

As the population continues to grow in the North and increasing use is made of resources, the consumptive use of water increases. Two principal concerns in this area include municipal water disposal and the storage of mining wastes from industrial processing. Other potential uses of the natural water environment that are of specific concern include hydro-power generation and large-scale diversions. These are considered below.

1.5.2.1. Hydro-Power

The total hydro-electric power potential north of 60°N is currently estimated to be 27,100 MW (16,100 within the N.W.T., Vassilev, 1990 personal communication; Northwest Territories Power Corporation; 11,000 MW within the Yukon, Monenco Consultants Pacific Ltd., 1983). The present operating capacity, however, is less than 0.5% of the potential (total of 126.6 MW; 79 in the Yukon and 47.6 MW in the N.W.T.; Simpson, 1989). Although small-scale (100-5,000 KW) hydro-electric units are under consideration for use by isolated communities in the North (Rueggeberg, 1985), the greatest hydro-electric potential is contained on the large rivers and their tributaries, particularly the Yukon and Mackenzie Rivers. For example, of the estimated 16,100 MW potential in the N.W.T., approximately 245 MW is from proposed sites under 50 MW but the remaining 98% is from larger capacity sites. Developmental pressure to use these larger-capacity sites could come from a desire to lure industry to the North with the promise of cheap power (Rueggeberg, 1985) or to export such power to southern locations. Whatever the case, potential impacts from such developments are likely to be wide-ranging and long-term. One example of the scale of

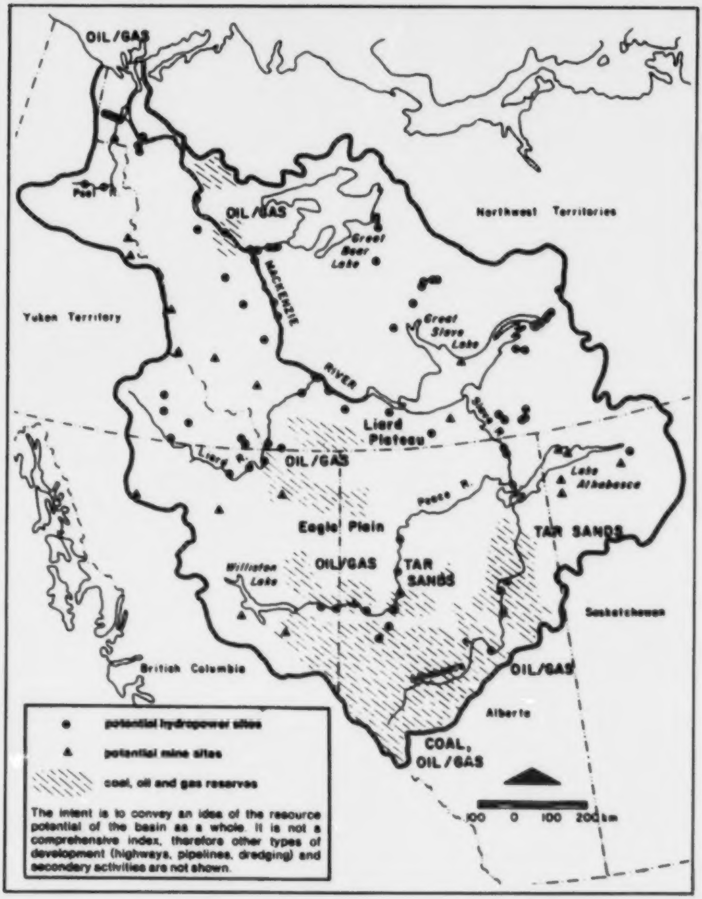
OIL/GAS
Peel R.
OIL/GAS
Great Bear Lake
Northwest Territories
MACKENZIE RIVER
Yukon Territory
Great Slave Lake
Slave R.
Liard Plateau
Liard R.
OIL/GAS
Lake Athabasca
Peace R.
Eagle Plain
TAR SANDS
Williston Lake
OIL/GAS
TAR SANDS
Saskatchewan
British Columbia
OIL/GAS
Alberta
COAL, OIL/GAS
100 0 100 200 km
● potential hydropower sites
▲ potential mine sites
▨ coal, oil and gas reserves
The intent is to convey an idea of the resource potential of the basin as a whole. It is not a comprehensive index, therefore other types of development (highways, pipelines, dredging) and secondary activities are not shown.

Figure 1.22. Selected development opportunties in the Mackenzie River basin *(after James and Foerstel, 1983).*

impacts from such developments can be gained from the host of environmental-impact studies conducted or identified by British Columbia Hydro in their evaluation of the upper Liard River for hydro-power construction (British Columbia Hydro, 1985). General subjects included: flow regime, sediment transport and channel morphology, ice regime, water temperatures and various related environmental and socio-economic impacts. Even potential modifications to the local climate of the Mackenzie Delta, over 1600 km downstream, were deemed important enough for scientific assessment (Harry and Parent, 1980).

1.5.2.2. Diversions

Small-scale water diversions have already occurred in the Territories, particularly in the Yukon as part of hydro-electric schemes (Rueggeberg, 1985). Some diversions in the headwaters of northward-flowing rivers are also being considered, such as an interbasin transfer from the Peace-Athabasca to the Saskatchewan-Nelson (Sadler, 1983). These diversions, however, are small compared to some of the mega-projects proposed to tap the major water resources of the North for export to drier southern regions. At least six major schemes have been proposed to divert northern rivers southward: 1) the North American Water and Power Alliance (NAWAPA), 2) the Central North American Water Project (CeNAWP), 3) the Kuiper Diversion Scheme; 4) the Western States Water Augmentation Concept, 5) the Magnum Diversion Scheme, and 6) the Great Recycling and Northern Diversion (GRAND) Canal. To provide some appreciation for the scale of these proposed diversions, the volume of water diverted in the NAWAPA scheme could be as much as 310×10^9 $m^3 \cdot yr^{-1}$, a volume comparable to the flow of the St. Lawrence River (Holm, 1988). The vision for many of these schemes is founded on the idea that the water carried by northward flowing rivers is largely being wasted. Although the broad environmental importance of this water has been established (e.g., Milko, 1987), such schemes still receive considerable attention and interest. Futhermore, while some feel that the large scale and huge costs of such schemes make their development highly improbable, it is worthy to note that Canada is already the international leader in diversions with more than 600 dams and 60 large domestic interbasin diversions. If combined, the total flow in these diversions would form Canada's third largest river, behind the St. Lawrence and the Mackenzie (Clark and Gamble, 1988). The issue of diversions of northern waters is likely to become increasingly topical in view of the growing need for water in the south.

1.5.2.3. Mining

Mining activities form almost 50% of the total economic base in the N.W.T. and 40% in the Yukon (1986 figures; Simpson, 1989). Well distributed throughout the North, they also include the world's most northerly mine on Little Cornwallis Island at latitude 75°23'N.

Mining and the related processing operations are the largest creators of industrial waste in the North. Often integral to these operations is the use of toxic chemicals. These and other chemical processes can produce noxious compounds and/or lead to the leaching of additional toxic substances from the mined material. Treatment and disposal of mining wastes is especially problematic in the North because the cold environment slows many of the degradation processes.

In most mining operations, the liquid wastes are retained in tailing ponds, either natural or man-made enclosures. Any discharge from the ponds must meet a host of environmental restrictions, but the largest problem is what to do with the liquid tailings (and solid wastes) when mining operations cease or when the pond capacity is exhausted (see Section 10.8.1.). Since, our limited knowledge about northern water balances precludes sound estimates of long-term pond capacity, wide margins of safety are necessary to avoid any potential overtopping and/or spilling of wastes into the local environment following pond abandonment. Storage and release problems are complicated by the presence of permafrost and the lack of knowledge about the related ground-water flow regimes. Little is known, for example, of the long-term effects of "warm" wastes on the retention capacity of permafrost-encapsulated ponds.

The other major mining activity affecting northern water resources is placer gold mining. Although widespread over the Yukon, most gold has been extracted from unglacierized portions of the Klondike, Sixtymile, Lower Stewart and Indian River drainages, lying primarily in the Yukon Plateau (Environmental Sciences Limited, 1990). Over the last five years, over 300 placer-mining licences have been issued by the Department of Indian Affairs and Northern Development (Milburn pers. comm., 1990). Major effects on the water resource include: the disturbance created by the extraction and sorting of the alluvial bed and bank material, the subsequent increase in downstream sediment loads, and the associated impacts on fish spawning and feeding areas (e.g., Envirocon Limited, 1986; Canada, 1989).

1.5.2.4. Oil and Gas

Oil and gas development continues to form a significant part of the northern economy (approximately 10% of the total economic base of the N.W.T.; Simpson, 1989). Although most production has centred on the northern mainland, it is now shifting to sites offshore and on the Arctic Islands. As discussed in Section 1.5.1., numerous hydrological studies were conducted in preparation for oil and gas development, focusing either on the local effects of the pipeline on adjacent permafrost or on specific hydrological hazards created by routing of the pipeline including, for example: accelerated erosion at channel crossings, disruptions created by icings and glacier outburst floods, scour and flooding produced by river-ice break-up, etc. One area of continued concern is the inability to model oil behaviour and transport in inland waters, especially under cold conditions where ice might be especially important.

1.6. EVOLVING ISSUES

1.6.1. Northern Pollution

Most developmental and industrial activities have a higher potential for producing significant pollution in the North than in more temperate regions because of environmental sensitivity associated with freeze-thaw processes and the temperature-dependent rates of chemical and biological processes that degrade pollutants. In view of the fragility and low assimilative capacity of northern ecosystems, cumulative and/or synergistic effects could be especially important.

Much pollution in the North originates from southerly latitudes and is carried northward by ocean currents, large river systems or the atmosphere. Of recent concern is the high level of contaminants found in various arctic mammals and in the Inuit diet. Research is being focused on this problem, but the precise pathways for many pollutants, particularly organic contaminants, remain poorly understood. Although few measurements have been made of organic contaminants in air, ocean currents or river run-off, PCBs and various chlorinated pesticides have been found in snow, ice, air and water in the North (Canada, Indian and Northern Affairs Canada, 1989).

One specific example of the airborne transport of pollutants into the North is the arctic haze. Aerosols originating in the western U.S.S.R. and Europe are transported in late winter to the Arctic of North America in large pancake-shape masses held together in the cold lower and middle atmosphere (Roots, 1990). Even if the concentration of airborne pollutants is low, the long duration of snow-cover in the North may result in the accumulation of significant quantities and a potentially large pollution shock during spring snowmelt. Details of specific pollutants are discussed in Section 7.3.

1.6.2. Global Change Effects

The effect of global change on the northern regions remains largely speculative partly because of the difficulty in dealing with surface-atmosphere exchanges in the presence of extensive ice and snow surfaces. Recognizing these problems, a plausible middle-of-the-road scenario, accepted by the 2nd International Workshop on Climate Change in Villach Austria - 1987, suggests that the average rise in winter temperatures north of 60°N will be approximately 2 to 2.4 times the global average, resulting in temperature increases of 6° to 7°C at about 70°N (Roots, 1990). Most warming is likely to occur in continental or coastal areas and be most pronounced during the winter period. In some extreme predictions, as much as a 15°C increase have been forecast to occur during the colder months but only a 1°C increase in summer. At lower latitudes, warming will be distributed throughout the troposphere but in the Arctic, where only very slight low-level mixing occurs, warming will be more confined, to a shallow, surface layer especially in winter under the stabilizing effect of arctic air masses (Harvey, 1982). The zones of highest temperature increase should also be at the snow to ice/snow boundaries where the albedo dramatically increases from low to high.

1.6.2.1. Precipitation/Snow/Glaciers

Although evaporation rates may not experience a significant redistribution over the globe (Harvey, 1982; see Section 8.4. for discussion of specific northern conditions), there should be a large-scale increase in the latent-heat transfer with the northward movement of snow and ice and hence, a northward shift of areas of maximum precipitation (Manabe and Wetherald, 1980). The combined effect of warmer air temperatures and an increase in the local surface-moisture supply (accruing from a decrease in ice cover) could increase precipitation along the arctic mainland coast and archipelago up to five-fold (Roots, 1990). Enhanced orographic activity should also increase precipitation on the windward slopes of the alpine zones (Harvey, 1982) resulting in a growth of the accumulation zones of glaciers common to these locations.

The precise hydrological response to be expected from the glaciers, however, remains somewhat in doubt and largely depends on the hypsometric distribution and response time of individual glaciers relative to the magnitude and rate of warming. If rapid melt of the northern glaciers and icefields occurs, sea level could rise by 0.5 to 1.5 m based on the Villach scenario (Roots, 1990). A rise of this order will be most significant for low-lying areas particularly the Mackenzie Delta where a large inundation could significantly alter the hydrological and biological system that links such a vast portion of the Canadian northland with the Arctic Ocean.

The impact of global change on snow in the North is complicated by the combined effects of temperature and precipitation. One likely scenario is that because of increased winter precipitation, the total winter snow accumulation will increase but, because of overall warmer temperatures, the duration of the snow season will be shortened (Barry, 1986). In general, an enhanced snow accumulation should lead to greater run-off and a higher potential for spring flooding. The total area impacted by such floods may, however, decrease marginally because of a northward shift of the snow and ice boundary that is estimated by Mokhov (1984) to be in the order of 2 ± 0.5 degrees of latitude per °C of warming.

1.6.2.2. Floating Ice

As for the snow season, the duration of river and lake ice should shorten giving the added benefit of a longer navigation period. Based on one model scenario, Barry (1986) notes the ice season will be shortened by about 15-20 days in the fall and 15 days in the spring for most of Canada. Warmer winter temperatures, however, will also lead to an overall decrease in ice thickness which will preclude surface-ice travel and transport in some regions. Thinner ice covers also could serve to decrease the higher spring-flood potential created by enhanced snow-melt run-off since many floods on northern rivers are ice-controlled (Section 5.3.1.).

1.6.2.3. Permafrost

Many long-term changes are likely to occur in the composition and distribution of permafrost, but there may be a considerable lag in their development because of the thermal inertia of the ice-rich materials (Smith, 1986). Warming within the continuous permafrost zone should result in an overall thinning of the permafrost and a northward retreat of the southern boundary. Mackay (1975), for example, speculated that it was located as far south as Fort Norman, N.W.T., 100 years ago, thereby giving a northward retreat rate to today's limit of approximately 3 $km \cdot yr^{-1}$. A number of modifications to the hydrological regime will develop, including: deepening of the active layer; expansion/contraction of thaw lakes (depending on the water balance); increased thermokarst activity; loss of lake and river bank stability and consequent modifications to drainage patterns and increased sediment loads; and greater discharge/recharge of the ground-water system. Some of these effects will be most pronounced within the discontinuous zone where a considerable reduction in the spatial coverage of frozen ground could occur especially where the layer is particularly thin. Northward migration of the permafrost zone could be especially important to northern wetlands that have developed from the ponding of water over existing frozen strata. At present, wetlands are the largest single contributor of methane, one of the greenhouse gases, to our atmosphere (Schindler and Bayley, 1990). If global change leads to wetter conditions, more methane will be discharged. However, if the melting of permafrost leads to a draining and a drying of these wetlands, their large storehouses of organic material will oxidize, thereby producing carbon dioxide, another greenhouse gas.

1.6.2.4. Other

A number of other water-related systems, particularly the distinct flora and fauna of northern ecosystems, may also be affected by global change. For example, experience in temperate latitudes has shown that common tree species and some grain crops are capable of migrating northward by approximately 100 km for every degree of warming, provided that sufficient moisture is available (Roots, 1990). The same has not been substantiated for northern species, but there is general consensus that some important long-term changes will occur in the northern ecosystems and that these may well have important feedback impacts on the northern hydrological regime.

Although some of the changes resulting from global change will have negative impacts, some positive effects on water-resources are also anticipated. In the case of Canadian sport fish, for example, the most important species thrive best in regions where air temperatures are between -1 and +2°C (Schindler and Bayley, 1990). At present the commercial fishery on Great Slave Lake rarely achieves the permitted annual limit of 1.7×10^6 kg (Simpson, 1989). With global warming of the northern latitudes, such large freshwater resources could develop appreciable stocks resulting in an expansion of the commercial fishery (Harvey, 1982). Similar benefits may well accrue to other water-related resource activities.

1.6.3. Monitoring Global Change

Considering the above potential impacts, the North is likely an excellent bellwether of global change. Water resources, in particular, offer a number of possibilities for interpreting and monitoring the rate and magnitude of global change. For example, one of the most successful attempts to reconstruct past climates has been achieved through analysis of the chemistry and pollen found within high-latitude glaciers; similar information can be found in alluvial deposits, particularly those in the deeper portions of larger lakes. Data concerning past temperatures and precipitation can be derived from records of glacial activity and a time-series of surface temperatures is preserved in the thermal record of permafrost. Additional opportunities for monitoring global change are offered by other water resources including, for example, the records of freeze-up and break-up of freshwater bodies and the mass-balance of glaciers.

1.7. REFLECTIONS

Given the magnitude and importance of Canada's northern water resources, surprisingly few researchers have specialized in northern hydrology. Moreover, because of the relatively unique character of the many scientific fields that comprise northern hydrology, researchers have often worked in relative isolation and developed narrow fields of specialization. Unfortunately, because of a poor understanding of the complexities of the North by a largely southern population, these northern researchers are frequently, and unfairly, expected to act as general experts on the North. This book will hopefully help to counteract this problem through the development of a broader understanding of the North's complex hydrological environment and lead to enhanced integration of the various fields in northern hydrology. Before delving into the specific details of northern hydrological research, the reader is asked to reflect on the following four quotes from a recent report dealing with polar science in Canada (Adams *et al.*, 1987).

> "Canada's science in the North ... tends to be old-fashioned. New techniques and new scientific theories spread slowly through Canadian northern research, mainly for economic and institutional reasons. This situation is not helped by a funding situation that discourages or even prevents the entry of bright young scientists into the world of Arctic research." (p.46)
>
> "The direct and indirect support by the federal government of polar science has declined in net terms since the early 1970's. During this period the federal government appears to have re-directed much of its support from long-term, sustained research and systematic data-gathering to the resolution of urgent short-term policy and political problems associated with non-renewable resource development or surveillance for defence." (p.vi)
>
> "To a distressing extent, research managers in government are more accountable for the "dollar output" of northern research than they are for research that is done in other parts of the country, mainly because of the proportionately higher costs of northern studies and the fact that a high proportion of northern research is funded by policy-related agencies or for industrial purposes. This situation discourages the scientific risk-taking that is essential to creativity. For many of the same reasons, university researchers are inclined or constrained to undertake studies in the North which tend to be scientifically safe and predictable and thus are less likely to be truly imaginative and intellectually daring." (p.46)
>
> "In the past, most science and research in the Canadian North has been undertaken or sponsored in a spirit of discovery and exploration, to address challenges of resource development or for military reasons. In future years, the issues driving Canadian needs for knowledge in the polar regions will be: enhancement of sovereignty; national security; social and economic pressures on rapidly growing northern population; arctic and global environmental quality; changes in wildlife habitat; the assumption of greater political authority by northeners; the concerns of non-northern public interest groups; and the need to have scientific knowledge from polar regions to address national and global problems." (p.vii) ❄

2

SNOW HYDROLOGY

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2.1. INTRODUCTION

Although snowfall is of small magnitude in the North, with some arctic areas referred to as "polar deserts", it can account for up to 80% of annual precipitation and blankets most of the northern landscape for up to half the year (Section 1.3.2.1.; Figure 1.8). This long-term storage of a significant portion of the annual precipitation only becomes available for the replenishment of liquid-water storage or as run-off during the brief spring melt period when air temperatures finally rise above 0°C. The resulting spring freshet plays an important role in the formation of ice jams and in the removal of ice from rivers and lakes (Section 5.2.), and often dominates the annual hydrograph both in terms of peak discharge and total volume of run-off (Church, 1974). As discussed by Church (1974), "the effect of snow storage, is ... the single most important feature of the hydrologic cycle everywhere in the North".

Snow plays other important, non-hydrological, roles in the northern environment. For example, the snowcover insulates the ground from the extreme cold typical of northern winters, resulting in more moderate ground temperatures beneath deeper snowcovers (Goodrich, 1982). As a result, the occurrence of permafrost is often controlled by the snowcover depth in the zone of discontinuous permafrost (Granberg, 1988). The snowcover is also ecologically important due to its effect on light transmission to underlying bodies of water (Adams, 1981), its insulating properties which result in a more favourable environment for animals and plants (Pruitt, 1970; Marchand, 1987), and its effect on feeding patterns of larger animals, for example Peary caribou (Miller *et al*., 1982). The occurrence of a snowcover has important implications to over-snow transportation techniques and to the building of snow roads (Adam, 1981). It also has a significant effect on the surface albedo (Male and Gray, 1981) and therefore the energy balance. Worldwide, snowcover increases the earth's albedo, resulting in lower global temperatures due to the snow-albedo-temperature feedback processes (Kukla and Kukla, 1974; Harvey, 1988).

This chapter emphasizes the processes controlling, geographical variations in, historical development of, and future research on the movement of water through snow and the resulting streamflow. It will also discuss briefly other aspects of snow hydrology; including the formation of snow in the atmosphere, and the growth and metamorphism of the snowcover during the winter and spring period. In addition, processes controlling the inclusion of solutes and pollutants into the snow, their redistribution within, and release from the snowcover will be reviewed.

2.1.1. Basic Processes

2.1.1.1. Atmospheric Processes

The formation of snow in the atmosphere depends on the availability of moisture and a number of processes which convert that moisture into particles with a large enough fall velocity to reach the ground. Snow formation begins as air cools due to uplift (either orographic or convective) or radiation cooling. When cooling occurs, the water vapour capacity of the air decreases, and with sufficient cooling, the air becomes supersaturated. However, condensation of water vapour to form liquid cloud droplets only occurs if condensation nuclei are present. These condensation nuclei are extremely common and very small in size, with typical concentrations and dimensions of 10^3 cm^{-3} and 0.1 μm respectively. Blowing dust is the most common source. The resulting cloud droplets, which have typical dimensions of 10 μm, will not automatically freeze at 0°C but may be supercooled as low as -40°C. In the presence of freezing nuclei, however, freezing of the droplets may begin anywhere between -3°C and -20°C. These freezing nuclei are not as common as condensation nuclei, and in fact only 1 in 109 aerosols are active freezing nuclei at -10°C. The process of nucleation by freezing nuclei is termed heterogeneous nucleation. At temperatures below -33°C droplets begin to freeze spontaneously and by -40°C all droplets freeze rapidly. This process is termed homogeneous nucleation. The importance of condensation and freezing nuclei is illustrated by the fact that 85% of snow crystals from the ice cap of northern Greenland have a clay particle at their centre and more than half of these are kaolin (Hobbs, 1973).

Clouds with temperatures of less than -40°C are composed of both ice particles and supercooled water droplets. Since the saturation vapour pressure with respect to ice is less than to liquid water (Neiburger *et al.*, 1973) there is a net transfer of water molecules from droplets to the air and then to the ice crystals. This growth of ice crystals at the expense of droplets is often termed the three-phase or Bergeron process. As the ice crystals grow, they begin to fall relative to the droplets; their growth rate accelerating due to both collision and coalescence with other droplets. Collision with supercooled droplets is often referred to as riming.

The initial shape of the ice crystals produced by the above processes is highly variable and is controlled by both the air temperature and supersaturation relative to ice (Magono and Lee, 1966; Nakaya, 1954). At various temperature and supersaturations, the predominant growth occurs on either the *c*-axis or *a*-axis of the ice crystals, resulting in crystals which have the form of plates, needles, hollow prisms, sector plates, or dendrites.

The formation of snow particles removes not only water vapour from the atmosphere, but also a wide range of other material as well, including aerosols, ions, trace gases, and pollutants. The two most important processes responsible for this removal include: (1) the incorporation of particles into the snow crystals within the cloud, which is often called "snowout", and (2) the scavenging of particulates below the cloud base, normally referred to as "washout". Variations in the efficiencies of these scavenging processes have received considerable attention in the literature (for example, see Pruppacher *et al.*, 1983). It is often hypothesized that snow particles have a higher collection efficiency than water droplets, resulting in higher concentrations of aerosols and ions in snow compared to rain (Anlauf *et al.*, 1986; Raynor and Hayes, 1983; Podzimek, 1987). An important factor controlling the efficiency of snow scavenging is the role played by riming (Borys *et al.*, 1983). Quantitative estimates of the scavenging efficiencies of snow have been difficult to obtain due to the complicated shapes of snow crystals (Wang and Sauter, 1985).

In addition to the process described above which incorporates impurities into or onto the snow particles during snowfall events, material may also be removed from the atmosphere by dry deposition processes. The relative importance of these "wet" and "dry" processes varies greatly, with "wet" processes dominating in regions of moderate to high precipitation (Dewalle, 1987). On a continental scale however, it is believed that wet and dry deposition are approximately equal in North America (Voldner and Sirois, 1986).

2.1.1.2. Snowcover Formation

In most areas of northern Canada, winter temperatures are almost continuously below 0°C, with above 0°C temperatures not occurring until the brief spring melt period. As a result, snowcover formation and development is dominated by dry-snow metamorphism, while wet-snow processes are restricted to the spring melt period. This section will deal with the dry-snow processes; wet-snow processes will be dealt with in the snowmelt run-off section.

In sheltered locations (e.g., forested sites), the snowcover is layered, with each layer representing a single snowfall event (U.S. Army, 1956), and its initial properties determined by the atmospheric conditions at the time of snow formation. The resulting snow depths are relatively uniform, but vary with vegetation type (Adams *et al.*, 1966) and with the proximity to trees (Woo and Steer, 1986). At exposed, windy sites (for example, tundra sites), however, the growth of the snowcover is dominated by wind drift events, and characterized by complex patterns of deposition and erosion. These snowcovers are layered, with each stratum representing a single drift event. They are highly variable in depth, with depth related to terrain type (Woo *et al.*, 1983a). The processes controlling blowing-snow erosion and deposition have been discussed by Schmidt (1982), and Pomeroy and Male (1987). The development of a snowcover dominated by blowing snow was described by Longley (1960). Blowing snow also affects the concentration and distribution of impurities on the snow grains (Pomeroy *et al.*, 1990).

Once the snow is deposited, its properties, in terms of grain size, shape, density, strength, and chemical composition, change so rapidly due to vapour flux induced metamorphic processes that the snow bears little resemblance to its initial form within a few days. In dry snow, two processes control these vapour fluxes. First, the saturation vapour pressure over the snow grain surface is controlled by the curvature of the surface, with higher vapour pressure over convex zones and lower over concave surfaces. The result is a vapour flux from convex to concave areas, and ultimately a rounding of the crystal (Colbeck, 1986a). This process results in slow changes to the crystal size and shape. Temperature gradients in the snowcover also help to establish the vapour gradients. Since these temperature/vapour gradients can be very large, rapid changes can occur to the snow crystals, with the rate of crystal growth and the crystal shape controlled by the magnitude of the gradients. With low gradients, a rounded snow crystal develops (termed equilibrium form), while with higher gradients a faceted snow crystal (or kinetic form) appears. The transition between equilibrium and kinetic forms occurs at approximately 10°C/m, although the exact transition is also dependent on snow density. Since the temperature gradient within a snowcover is constantly changing in response to air temperature, it is common to find grains in various stages of the equilibrium and kinetic forms, and with mixed equilibrium/kinetic characteristics.

Equilibrium grains, typically between 0.1 to 0.5 mm in diameter, are relatively strong due to the development of necks (sintering) between the individual grains. Kinetic grains are larger in size, with diameters of up to 5 mm. The shape of these grains varies from hexagonal prisms at lower growth rates to hollow, depth-hoar grains at the highest growth rates. Since sintering does not occur at these higher growth rates, kinetic grains are of low strength.

In addition to the above processes which occur within the snowcover, a number of surface processes occur which produce distinct snow grains or crusts. These surface crystals, which may be buried by a subsequent snowfall, include hoar, wind crust, melt-freeze layers, sun crusts and freezing-rain crusts. The final result of the entire suite of deposition and metamorphic processes is a snowcover composed of numerous layers, each with different properties such as grain size, porosity, strength, density, and permeability, with each layer possibly separated by ice crusts or thin layers of low permeability snow. This complex, layered snowcover, which has often been assumed to be homogeneous, has important implications to snowmelt run-off.

2.1.1.3. Snowmelt Run-Off Processes

2.1.1.3.1. Energy Balance

As described in more detail in Section 8.1.1., the snowcover energy balance can be written as:

$$Q_M = Q^* + Q_H + Q_E + Q_R + Q_G - du/dt \qquad [2.1]$$

where Q_M is the energy available for melt, Q^* is net all-wave radiation, Q_H and Q_E are the sensible- and latent-heat fluxes, Q_R is the flux of heat from rain, Q_G is ground heat flux at the snow-soil interface, and du/dt is the rate of change of internal energy. Numerous studies have described methodologies for estimating the value of Q_M at a site from temperature index methods (Male and Gray, 1981), or by determining the magnitude of the individual terms in Equation 2.1 and therefore the value of Q_M, from the bulk transfer approach (Moore, 1983), eddy-correlation methods

(McKay and Thurtell, 1978), or profile methods (de la Casinière, 1974). In temperate areas, Q_G and du/dt are normally of small magnitude. As a result Q_G is often assumed to be zero, while du/dt is simply treated as the cold content of the pack. In cold northern regions, however, they are often larger in magnitude and therefore must be treated directly (Marsh and Woo, 1987). In addition, sublimation (Q_E) may be large sometimes, significantly reducing the water available for run-off. Such variations in melt may only be predicted from detailed energy-balance criteria. Although temperature index methods work well in "normal" conditions, they often fail in extreme or unusual circumstances.

2.1.1.3.2. Wetting Fronts and Ice Layer Growth

When surface melt first begins, a wetting front develops in the snowcover, producing an upper layer of wet snow isothermal at 0°C and a lower layer of dry snow with a temperature below 0°C (Yosida, 1973). The rate at which this horizontal, one-dimensional wetting front ($d\xi/dt$) moves downward is given by (Colbeck, 1976):

$$d\xi/dt = U / \{\phi S_w + (\phi - 1)[(T_s \rho_i)/(L_f/C_i) \rho_w]\} \qquad [2.2]$$

where U and S_w are the vertical, non-saturated meltwater flux and water saturation immediately above the wetting front, ϕ is the snow porosity, T_s is the snow temperature immediately below the wetting front, ρ_i and ρ_w are the ice and water densities respectively, L_f is the latent heat of fusion, and C_i is the specific heat of ice.

Because of inhomogeneities in the porous media and flow instabilities, due to differences in the density of water above the wetting front and in the air below the front, wetting fronts in snow are not horizontal, but instead develop into complex, three-dimensional fronts. Seligman (1936), U.S. Army (1956), Wakahama (1968), and Marsh and Woo (1984a) have described wetting columns or fingers which are narrow in width and relatively evenly spaced, at the leading edge of wetting fronts in snow (Figure 2.1). Similar features have been reported by Hill and Parlange (1972) from laboratory studies of wetting fronts in sand. Wankiewicz (1979) described a conceptual model of wetting fronts in stratified snowcovers. His FINA model (Flow Impeding, Neutral, or Accelerating) proposes that the effect of a horizon on a wetting front is dependent on the pressure on either side of the horizon, with simple horizons either: (1) impeding the flow with water accumulating above the horizon, then flowing laterally down-slope until a vertical stream penetrates the horizon at some point down-slope, (2) accelerating the flow if the wetting front becomes unstable and fingering occurs, due to infiltration instability with horizontal variability acting only as a trigger, or (3) having no effect on the flow. In this situation the horizon is neutral and the wetting front passes it unchanged. Wankiewicz (1979) also suggested that the flow pattern at horizons could be complicated by compound horizons. For example, a thin fine-grained layer between two coarse-grained layers could impede the flow and then accelerate the flow beneath the layer. Unfortunately there have been no systematic studies of these processes in snow.

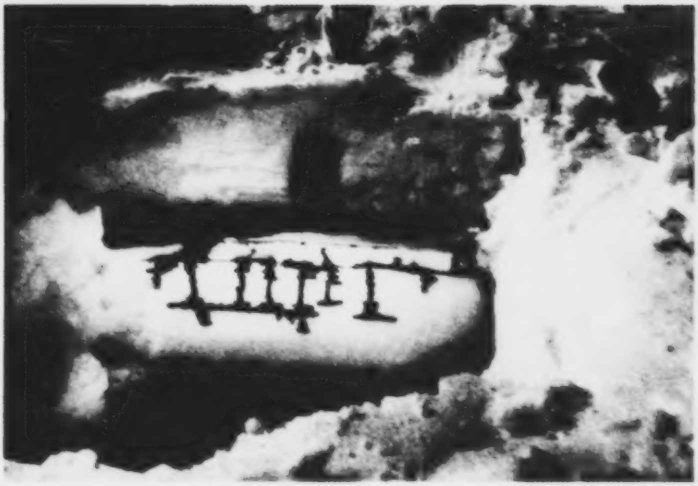

Figure 2.1. Flow fingers at the leading edge of wetting fronts in a layered snowcover. Note the development of a wet layer at premelt horizons and flow fingers developing below the horizons.

The development of flow fingers at wetting fronts results in the formation of three zones within a snowcover: (1) an upper zone of wet snow at 0°C, (2) a zone of mixed wet and dry snow, with wet areas at 0°C and dry areas with a temperature below 0°C, and (3) a lower zone of dry snow with a temperature of less than 0°C (Figure 2.2). In certain circumstances, "pockets" of dry snow may remain long after run-off has begun.

Since conduction cannot occur in the upper layer of isothermal wet snow and the underlying frozen soil can only remove energy from the snow, the only process that can warm the snow during the melt period is the freezing of meltwater within the snowcover. This freezing of meltwater releases latent heat which is then conducted away from the ice body and used to warm the surrounding dry snow and underlying ground. Re-freezing of meltwater occurs within the zone of mixed wet and dry snow at the leading edge of an advancing wetting front, within flow fingers to form vertical ice columns, at pre-melt stratigraphic layers to form horizontal ice layers, and at the base of the snowcover to form basal ice. In the early stage of development these ice bodies are composed of large polycrystalline melt-freeze particles (3 to 5 mm in diameter), which are frozen together into a permeable, honeycomb structure. If freezing continues, this honeycomb structure is transformed into low-permeability ice with a density of over 800 kg·m^{-3} (Marsh and Woo, 1984a).

Ice columns typically range in size from 20 to 90 mm in diameter, though massive ice columns have been observed in the snow and firn of cold glaciers (Müller, 1963). Ice layers are commonly from 1 mm to 200 mm in thickness, and vary in horizontal extent from less than 0.2 m for ice layers within

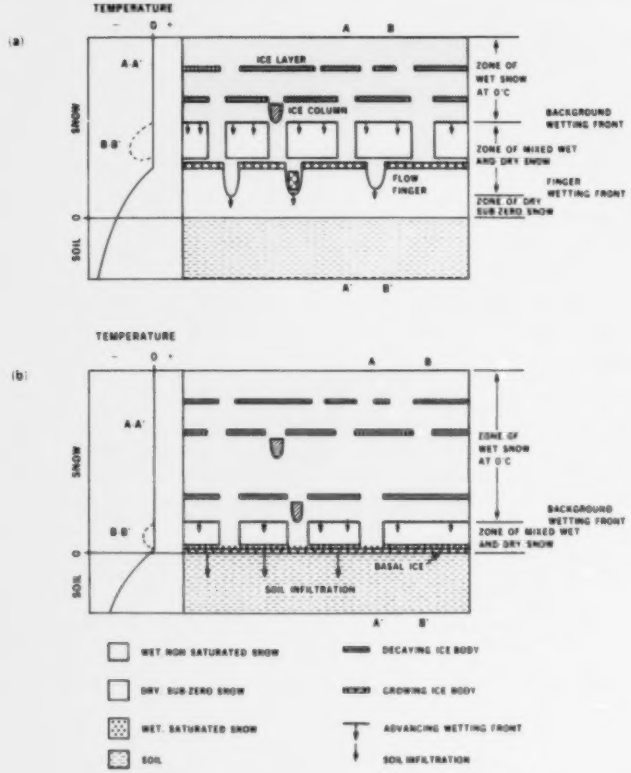
(a)
TEMPERATURE
- 0 +
A B
A-A'
B-B'
SNOW
SOIL
0
ICE LAYER
ICE COLUMN
FLOW FINGER
ZONE OF WET SNOW AT 0°C
BACKGROUND WETTING FRONT
ZONE OF MIXED WET AND DRY SNOW
FINGER WETTING FRONT
ZONE OF DRY SUB-ZERO SNOW
A' B'
(b)
TEMPERATURE
- 0 +
A B
A-A'
B-B'
SNOW
SOIL
0
ZONE OF WET SNOW AT 0°C
BACKGROUND WETTING FRONT
ZONE OF MIXED WET AND DRY SNOW
BASAL ICE
SOIL INFILTRATION
A' B'
WET NON SATURATED SNOW
DRY SUB-ZERO SNOW
WET SATURATED SNOW
SOIL
DECAYING ICE BODY
GROWING ICE BODY
ADVANCING WETTING FRONT
SOIL INFILTRATION

Figure 2.2. Hypothetical example showing water movement into initially cold, dry snow; (a) before meltwater reaches the snowcover base and (b) after meltwater reaches the base, but before the entire snowcover is wet *(from Marsh and Woo, 1984a).*

stratigraphic units, to more than 3 m for those at horizon boundaries separating distinct snow layers (Marsh and Woo, 1984a). Ice layers vary from highly permeable, for those formed by a minimum of freezing, to virtually impermeable for fully-developed layers. The permeability of these ice layers increases rapidly as their temperature rises above -0.1°C due to the increase in the size of the veins between crystals (Langham, 1974a, 1975). If the ice layer experiences changes in temperature above or below -0.1°C, the permeability of the ice layer will also vary (Langham, 1974b). As a result, ice layers may undergo diurnal changes in permeability.

An ice layer will also form at the base of the snowcover if the substrate has a low permeability, resulting in a saturated layer forming at the snow base, and the soil heat flux is negative (i.e., the soil is colder than the snow). This basal ice layer will continue to grow until the soil heat flux is reduced to zero or the snowcover is removed by melting. These basal ice layers have been reported on glaciers, where they are called superimposed ice (Wakahama *et al.*, 1976), in alpine regions (Gardner, 1983), and in permafrost environments (Woo *et al.*, 1982). Basal ice layers may grow to greater thicknesses (up to 0.71 m reported by Woo *et al.* (1982) for a seasonal snowcover) than ice layers within the snowcover since their growth period is normally of longer duration.

The total volume of water which re-freezes within and at the base of the snowcover is determined by the initial snow temperature and the soil heat flux. In temperate areas, the soil heat flux is negligible, so that re-freezing in the snow is controlled entirely by the cold content of the snow. When the snow is underlain by frozen soils, however, the soil heat flux can be large, with a resulting increase in the volume of water which freezes in the snow.

2.1.1.3.3. Grain Growth in Wet Snow

As soon as the snow is wetted, the snow grains begin to grow rapidly due to the effect of grain size on the equilibrium temperature among the solid, liquid, and vapour phases of water at the scale of individual grains. Since the equilibrium temperature increases with grain size, the resulting temperature gradient between particles of different radii produces a heat flux from the larger to the smaller particles. As a result, the larger grains grow at the expense of smaller grains, smaller grains disappear, the mean grain size increases with time, and the grains become rounded. Since the heat flux is predominantly through the liquid/solid interface, the heat flux and therefore growth rate is greater at higher water saturations.

At low water contents and in the absence of macroscale melt-freeze cycles, collections of individual, well-rounded grains form, called grain clusters. When melt-freeze occurs, the resulting particles often appear similar to the grain clusters, but they do not have the same distinct well-rounded appearance. With higher liquid-water contents, the grain clusters or melt/freeze particles break down into cohesionless grains.

Colbeck (1986b) suggested that the rate of grain growth in wet snow can be described by:

$$d = d_o + A\,t^B \qquad [2.3]$$

where d is the grain diameter, d_o is the initial grain diameter and t is the time since the snow was first wetted. Using laboratory data, Colbeck (1986b) found that coefficient A = 0.42 and coefficient B = 0.36 for saturated snow and Colbeck (1976) suggested that A = 0.1 and B = 0.5 predicted the growth rate for "expected water saturation following a heavy rainfall on cold snow". From field measurements of grain growth at Resolute Bay, Marsh (1987) found that A = 0.44 and B = 0.89. This gives a more linear growth rate than that suggested by the laboratory studies. Marsh (1987) suggested that this occurred because of the complicated sequence of metamorphic events that occur in natural snowcovers, including changes in water saturation and melt/freeze cycles. Such variations do not occur in laboratory studies.

This increase in grain size with time, results in a concomitant increase in snow permeability as suggested by the relationship between saturated permeability (κ_s) and grain size (Shimizu, 1970).

$$\kappa_s = 0.077\ d^2\ \exp(-7.8\ \rho_d/\rho_w) \qquad [2.4]$$

where ρ_d is the dry snow density and ρ_w is the density of water. Unlike most other porous media, therefore, the saturated permeability of snow does not remain constant, but increases with time. This phenomena has an important effect on snowmelt run-off, but is often ignored in modelling studies.

2.1.1.3.4. Vertical, Unsaturated Water Flux in Wet snow

Assuming normal melt and rainfall rates, and given the coarse grain size and high permeability of snow, the vertical flux of water through freely-draining snow is always non-saturated. The Darcy equation states that this flux of water is given by:

$$U = -\ (\kappa/\mu_w)(\partial p_w/\partial z + \rho_w\, g) \qquad [2.5]$$

where z is the elevation above a datum, κ is the permeability, μ_w is the dynamic viscosity of water, g is the acceleration due to gravity, and p_w is the water pressure.

The application of Equation 2.5 has generally followed either the gravity-flow solution derived by Colbeck (1972, 1978) or the application of numerical techniques (Jordan, 1983; Bengtsson, 1982; Illangasekare *et al.*, 1990). In the following section, Colbeck's gravity-flow solution will be presented since it illustrates a number of interesting aspects of water flux in snow. In certain circumstances though, it should be realized that the numerical techniques are superior.

Colbeck (1972, 1978) assumed that the water-pressure gradients are normally small compared to the gravitational gradients, and therefore Equation 2.5 simplifies to:

$$U = -\ (\kappa/\mu_w)\ \rho_w\, g \qquad [2.6]$$

Colbeck further assumed the permeability followed a power relationship between the saturated permeability (κ_s) and the effective water saturation (S^*) as $\kappa = \kappa_s \cdot S^{*\varepsilon}$ (where ε is a coefficient), so that Equation 2.6 may be written:

$$U = -\ \Lambda\ \kappa_s\ S^{*\varepsilon} \qquad [2.7]$$

where: $\Lambda = (\rho_w\ g)/(\mu_w)$. Using Equation 2.7 in the continuity equation, Colbeck showed that any value of flux U propagates through the snow at the speed $dz/dt\big|_U$ given by:

$$dz/dt\,\big|_U = (\varepsilon\ \Lambda^{1/\varepsilon}\ \kappa_s^{1/\varepsilon})/\phi^*\ U^{(\varepsilon-1)/\varepsilon} \qquad [2.8]$$

where ϕ^* is the effective porosity calculated as $\phi^* = \phi(1 - S_i)$, ϕ is the snow porosity and S_i is the irreducible water saturation.

Equation 2.8 demonstrates, that when the properties of the snowcover remain constant, the rate of movement of flux U increases with increasing melt-water flux. As a result, during diurnal snowmelt events slow-moving fluxes in the morning are overtaken by faster-moving afternoon fluxes. Assuming that pressure gradients are small, then a flow discontinuity or shock front develops which is bounded by large fluxes (U+) above the front and smaller fluxes (U-) below the front. The rate of propagation of this shock front ($d\zeta/dt$) can be calculated as (Colbeck 1978):

$$d\zeta/dt = \Lambda^{1/\varepsilon} \; \kappa_s^{1/\varepsilon}/\phi^* \left(U_+^{(1-\varepsilon)/\varepsilon} + U_+^{1/\varepsilon} U_-^{1/\varepsilon} + U_-^{(1-\varepsilon)/\varepsilon}\right) \qquad [2.9]$$

Equation 2.8 and 2.9 may be solved graphically (Dunne *et al.*, 1976) or numerically (Tucker and Colbeck, 1977) to predict water flux at any depth in the snowcover.

The application of these equations demonstrates the main features of meltwater flux waves in snow. With increasing depth, for example, the rise to the peak steepens, the peak flow decreases in volume and arrives later, and the recession limb of the hydrograph lengthens and the minimum flow increases (Figure 2.3).

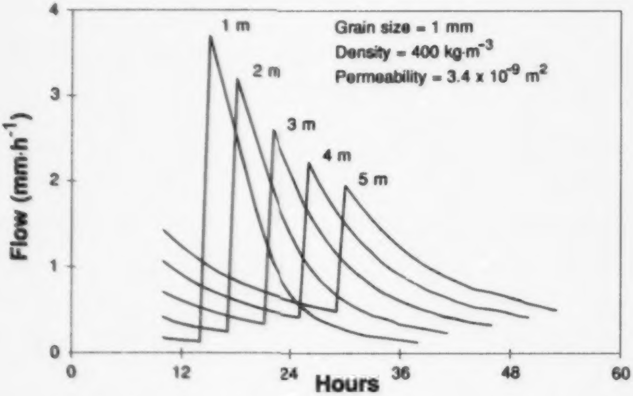

Figure 2.3. Hydrographs at increasing depths (from 1 to 5 m) within a snowcover as predicted using Equations 2.8 and 2.9. Note that with increasing depth, the rise to the peak steepens, the peak flow decreases in volume and arrives later, the recession limb of the hydrograph lengthens, and the minimum flow increases.

The most difficult parameters to determine appropriate values for in order to use Equations 2.8 and 2.9 are κ_s and ε. Typically, they are determined from drainage measurements of wave celerity during periods of declining flow (Marsh, 1990). The need to measure the flux in the snow in order to estimate values of κ_s and ε, before modelling the flow, is a major limitation to applying Equation 2.8 and 2.9.

Values of κ_s derived from drainage measurements of wave celerity are often termed equivalent permeability (κ_e) (McGurk and Kattelmann, 1986) since they represent integrated values over the entire snowpack depth. Values of κ_e are significantly different from κ_s, calculated from Equation

2.4, because of the effect of ice layers and flow fingers. It is interesting to note that whereas κ_s increases in value during melt metamorphism, κ_e decreases in value as the snowpack ripens. This is illustrated in Figure 2.4, showing the Shimizu relationship between density and grain size, and saturated permeability. Also shown in Figure 2.4, is an example of changes in κ_e as determined by McGurk and Kattelmann (1986) over a 28-day period of ripening.

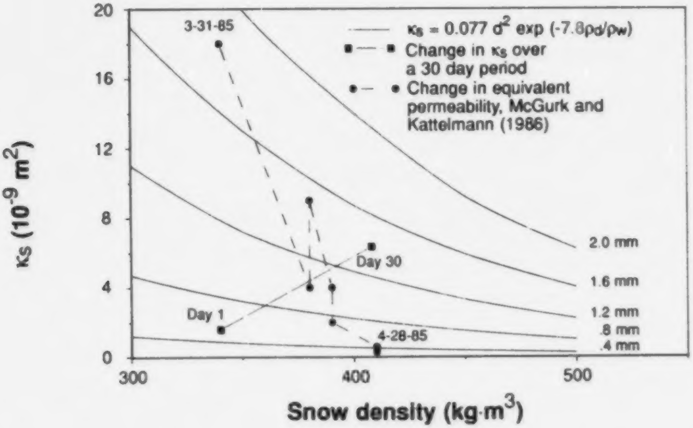

Figure 2.4. Relationship between snow density, grain size and saturated permeability (κ_s) calculated from Shimizu (1970) (Equation 2.4). Also shown are the change in equivalent permeability (κ_e) over a 30-day period in the spring of 1985, and changes in κ_s as predicted from Equation 2.4 assuming a change in grain size as predicted by Equation 2.3.

The decrease in permeability contrasts sharply with the predicted increase in κ_s predicted by grain growth (using Equation 2.3) and a measured change in snow density, and the Shimizu relationship (Equation 2.4). These contrasting observations can be reconciled as follows. At a small scale (i.e., tens of cm), κ_s increases as predicted by the Shimizu relationship (Equation 2.4) due to the grain size increase. At a larger scale, κ_e is controlled by κ_s and by flow fingers and ice layers. These flow fingers/ice layers concentrate water into a small area, thus accelerating the flux of water through the snow (Equation 2.8) and therefore increasing κ_e. As ripening progresses, the snow becomes more homogeneous and the fingers less dominant, and as a result κ_e decreases with time. The overall result is that in natural snowcovers meltwater can move more quickly through fresh wetted snow than older, melt-metamorphosed snow.

The second major unknown in Equation 2.8 and 2.9 is the value of ε. Reported values range from 1.5 to 4.8 (Marsh, 1990). Although ε would be expected to be related to snow properties, Denoth *et al.* (1979) stated that "no clear relationship exists between any snow parameter and ε". Given the lack of a method to predict ε, most studies have assumed that $\varepsilon = 3$ for all snowcover conditions.

2.1.1.3.5. Lateral, Saturated Flow at the Base of the Snowcover

When wetting fronts reach the base of the snowpack, meltwater may infiltrate the underlying soil, or run-off laterally in a saturated layer much like overland flow. Colbeck (1978) showed that when the ground slope (θ) is large compared to the "spatial gradient of the thickness of the saturated layer $\partial h/\partial x$", then flow through the saturated layer is given by:

$$\Lambda \kappa_s \theta (\partial h/\partial x) + \phi (\partial h/\partial t) = I \qquad [2.10]$$

where h is the thickness of the saturated layer, x is the co-ordinate along the base of the snowpack and I is the net influx to the saturated layer. Colbeck (1978) showed the speed of the water particles (c) in a saturated layer at the base of the snowcover is given by:

$$c = \Lambda \kappa_s \theta \phi^{-1} \qquad [2.11]$$

The discharge per unit width (q_L) of the saturated layer at time (t_L) is then:

$$q_L(L,t_L) = \Lambda \kappa_s \theta \phi^{-1} \int_{t_o}^{t_L} I(t)dt + q(0,t_o) \qquad [2.12]$$

where L is the length of the saturated layer, $q(0,t_o)$ is the lateral flow in the saturated layer at L=0, and the integration period (t_L-t_o) is given by:

$$(t_L-t_o) = \Lambda^{-1} \kappa_{s-1} \theta^{-1} \phi L \qquad [2.13]$$

2.1.1.3.6. Snowmelt Run-Off Lag Times

For a hillslope underlain by an impermeable substrate, the lag time between melt at the snow surface and outflow to a stream channel at the base of the hillslope, is controlled by both the lag time due to the vertical unsaturated percolation (t_u) and the lag time due to lateral saturated flow at the base of the pack (t_s). These lag times may be calculated as (Colbeck, 1978):

$$t_u = \phi^* D_s (3 \Lambda^{1/\varepsilon} \kappa^{1/\varepsilon} U^{(1-\varepsilon)/\varepsilon})^{-1} \qquad [2.14]$$

and

$$t_s = \phi L (\Lambda \kappa_s \theta)^{-1} \qquad [2.15]$$

where D_s is the snowpack depth.

For typical snowcover characteristics and melt rates, t_u is much larger that t_s as shown in Figure 2.5. For example, with a snow depth of 1 m and a slope length of less than 1000 m then $t_u > t_s$, and for slopes longer that 1000 m, $t_u < t_s$.

2.1.1.4. Snowcover Chemistry

Metamorphic processes not only affect the physical structure of the snowcover, but also the distribution of impurities within the snowcover and the release of solutes from the pack. The continuous evaporation/condensation

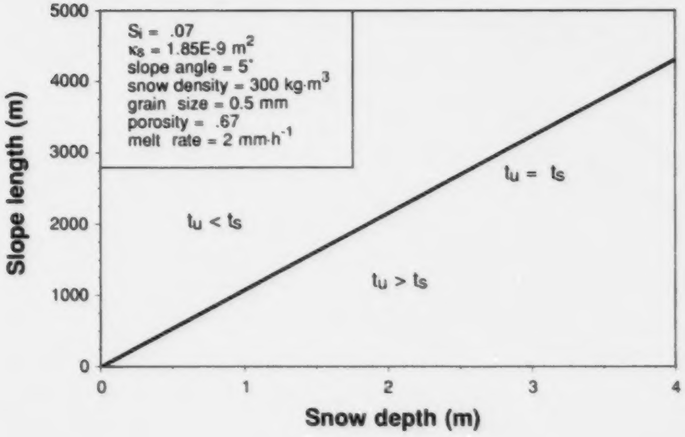

Figure 2.5. Comparison of the lag times due to vertical unsaturated flow (t_u) and horizontal saturated flow at the base of the snowcover (t_s) for given slope lengths and snow depths.

of water molecules in dry snow and melting/freezing in wet snow cause a segregation of impurities because the impurities are not readily incorporated into the crystal lattice during recrystallization (Colbeck, 1981). Although, the efficiency with which solutes are rejected from the crystal lattice is highly variable (Davies *et al.*, 1987), the overall result of this process is to purify the snow grains and concentrate impurities on grain surfaces (Johannessen and Henriksen, 1978; Colbeck, 1981).

Since the bulk of impurities are located on the outside surface of snow particles, the soluble components are readily available for removal from the snow particle and for incorporation into the first wave of meltwater moving through the snowcover. The available field and laboratory work clearly illustrates that this process results in the early components of the meltwater being enriched in soluble ions compared to the bulk composition of the snow (Johannessen and Henriksen, 1978). Due to the influence of melt/freeze within the pack, variations in melt rate, and the effect of flow fingers, the fractionation is highly variable from site-to-site and from year-to-year. In general though, field studies have found that up to 80% of the total solute load of a snowpack is removed in the first 20 to 30% of the melt quantity (Johannessen and Henriksen, 1978; Davies *et al.*, 1987) and that in this early stage of melt, the solute concentration is 2 to 5 times the initial concentration of the snowcover. In regions with high pollutant loadings, these processes can result in the sudden release of pollutants from the snow and therefore rapid changes in the chemical composition of lakes and streams (Johannessen and Henriksen, 1978). Chapter 7 details some of the snow-related pollutants affecting the North.

In addition to the above processes, it is well known that preferential elution occurs, such that certain ions are released earlier than others. Unfortunately there is little laboratory information on the microscale details of this fractionation process (Davies *et al.*, 1987) and there is insufficient data to enable prediction of this phenomena.

2.1.2. Geographical Variations

2.1.2.1. Snowfall

Snowfall amounts are controlled by (1) the pattern of weather systems, (2) the presence of mountains, (3) air temperature, with colder air retaining less moisture, and (4) the amount of moisture in the air (precipitable water), related to both local and distant sources of water. Local sources, including evaporation from land and local bodies of water, is highest in the summer, gradually decreasing in amount as the land becomes snow-covered and lakes and oceans freeze. In the Canadian North, important distant sources of water include the North Pacific, the Arctic Ocean, and Baffin Bay. The influence of the Pacific is limited by the Western Cordillera, while that of the Arctic Ocean is limited by the extensive ice cover. As a result, both precipitable water and snowfall are low over much of northern Canada, decreasing in a northerly direction, and decreasing in magnitude over the winter as local sources of water are cut off.

Annual snowfall ranges from less than 0.8 m in the northwestern sections of the Arctic islands to over several metres in the St. Elias Mountains in the southwestern Yukon. When expressed as water equivalent, the proportion of the annual precipitation which falls as snow varies widely. For example, at high elevations in the St. Elias Mountains and in the Arctic Islands, snow can in some years represent 100% of annual precipitation, while at Alert at the northern tip of Ellesmere Island, it is 89% and at Whitehorse in the southern Yukon, it is 44% of annual precipitation. As will be discussed latter, these values for annual snowfall, and therefore the percentage of annual precipitation which falls as snow, must be viewed cautiously due to the severe undercatch problems.

The relative importance of various sources of water, and local versus distant sources and the regional variations in these, are not well known. Although it is beyond the scope of this chapter to provide a comprehensive review of snowfall source areas for northern Canada, it is instructive to look at a typical example, in this case the Queen Elizabeth Islands, since it illustrates some important points.

The extensive polynya in northern Baffin Bay (called the "North Water") has received considerable attention in terms of its influence on precipitation in the Queen Elizabeth Islands and on the mass balance of glaciers on Ellesmere Island, Devon Island, and Greenland. Studies of the O^{18} content of snow in the Queen Elizabeth Islands has provided information on the role of local (i.e., the North Water) versus distant sources of water. Using a simple, global oxygen isotope model, Fisher and Alt (1985) suggested that for coastal sites at a latitude of 75°N, approximately 25% of the precipitation should be of a local origin, with the remainder originating from between 20 and 65°N. This value is in agreement with those measured by

Schriber *et al.* (1977), who found that 25% of snowfall around the northern end of Baffin Bay was derived from the North Water. Similarly Koerner and Russell (1979) found that on the south-eastern side of Devon Island approximately 30% of the moisture was of local origin (i.e., Baffin Bay) but that it decreased to only 8% at the top of the ice cap. This rapid decrease in the percentage of local water, and the rapid decrease in snowfall with distance from Baffin Bay (Koerner, 1979), suggests that the influence of the North Water is limited to a small geographical area. For areas further away from Baffin Bay, snowfall sources are not well known. Studies of O^{18} content in snow by Koerner (1979) and Jeffries and Krouse (1987) suggested that the influence of the Arctic Ocean is small, and is limited to the western edge of the Queen Elizabeth Islands.

The large-scale air-mass movements which play an important role in controlling snowfall distribution, also play a crucial role in the chemistry of the snow and the snowpack. Barrie (1986) showed that the winter arctic haze phenomena is controlled primarily by the transportation of dust and pollutants from mid-latitude sources in Eurasia. The spatial and temporal distribution of these pollutants has been determined from analysis of snowfall samples collected at northern weather stations (Barrie and Hoff, 1985), from analysis of seasonal snowpack samples (Shewchuk, 1986), and from glaciers (Barrie *et al.*, 1985).

2.1.2.2. Snowcover

The snowcover begins to develop as early the end-of-August in northernmost Canada, or as late as late-October in more southerly areas of the Canadian North. Likewise, removal dates vary from as early as the middle-of-May to as late as late-June, giving a snowcover duration of approximately 190 to 300 days per year (Figure 1.8).

Since winter snowmelt periods are uncommon in the Canadian North, snowcover depth would be expected to be closely related to snowfall amounts. This is generally true in forested locations. In these environments the mean maximum snow depth varies from as little as 30 to as much as several metres (Figure 1.9). In tundra areas, however, it is more difficult to report a mean snow depth. Figure 1.9 reports values for the tundra areas, but since snow depth is highly variable (Woo *et al.*, 1983a) these numbers reflect only the snow conditions at the survey site. They are not representative of larger areas.

2.1.2.3. Nival Streamflow Regimes

The sequence of snow storage during the winter, followed by snowmelt in the spring, effectively redistributes six to ten months of precipitation into the brief snowmelt period. As described in Section 1.4.1. Church (1974) has classified the resulting streamflow as either subarctic-nival or arctic-nival, but as found by Marsh and Woo (1981) perennial snowbanks can extend the snowmelt period throughout the summer, and as a result, the annual hydrograph is very similar to that of a proglacial environment. Lewkowicz and Young (1990) also reported that run-off from a small basin dominated by a large perennial snowbank may be considerably larger than the annual precipitation and have a proglacial run-off regime.

2.1.3. Relationship to Other Hydrological Variables

Seasonal snowcovers play an important role in many aspects of the northern hydrological cycle. One of the most important is the role of snow as an efficient insulator. Since snow on the ground occurs only during the coldest months of the year, its net effect is to raise the mean annual ground temperatures by several degrees (Goodrich, 1982). Using a simulation model, Goodrich (1982) showed that in permafrost regions the mean annual ground temperature may be raised a few degrees due to the existence of a 0.5 m thick snowcover. In addition, Goodrich found that the autumn snow conditions are critical, but that the melt rate during the spring is not. Goodrich also noted that a major problem with predicting ground temperatures in permafrost regions is our inability to model the development of the snowcover and the changes in snow properties during the critical autumn period. At present, there are few models available for predicting snowcover properties during the winter. Anderson (1976) presented a point-energy mass-balance model, but it has received minimal testing in northern regions (e.g., Stein *et al.*, 1987). Probably the most extensive data sets on the relationship between ground temperatures, the occurrence of permafrost, and snowcover characteristics are those reported by Granberg (1988) for the Schefferville area.

Snowmelt run-off and permafrost hydrology are also intricately linked. During melt metamorphism, the freezing of water in and at the base of the snowcover provides an important source of energy which is used to warm the underlying soil. As a result, the active layer is often warmed to near freezing before the snowcover is removed. The snowcover also supplies considerable water to the active layer. Early in the melt period, meltwater may infiltrate into the frozen soil. The volume of water may be very large in dry, highly-porous soils with large macropores (Marsh, 1988) or very small in soils which were saturated at freeze-up (Marsh and Woo, 1984a). The portion of this water which re-freezes in the soil is dependent on the grain size and temperature of the soil.

In areas where snow depth is highly variable, the snowcover becomes discontinuous soon after the start of melt. In this situation, meltwater released from residual snowbanks enters into the active layer. An important but little-studied aspect of northern hydrology concerns the period during which the basin changes from snow dominated to active-layer dominated. The suitability of snowmelt models for application in permafrost environments is examined by Landine *et al.* (1988). Other modelling efforts are reviewed in Section 3.4.2.

2.2. HISTORICAL BACKGROUND

Few snow-hydrology studies were conducted in northern Canada prior to the mid to late 1950s. During the 1950s, however, road, airstrip and mining developments created a need for information on basic snowcover properties. Some of the early work on snow hydrology was conducted by large group projects such as the McGill University Sub-Arctic Research Station at Schefferville and the Jacobsen-McGill Arctic Research Expedition to Axel

Heiberg Island. However, most of the snow-hydrology work conducted in northern Canada has been performed by small university and government research programs.

2.2.1. Snowfall

A major research thrust during the last thirty years has been concerned with determining the actual amount of snowfall during the winter period. Jackson (1960b), for example, described the principal sources of error involved in snowfall measurements. He noted that in many more temperate areas snowfall estimates are about 80 to 90% of actual, but that "in high latitudes the error may be quantitatively greater" due to higher average wind speeds and the large number of trace events. By the late 1960s and early 1970s water balance studies were providing the first direct evidence of snow undercatch problems. Findlay (1969b) in the Schefferville area and Cogley (1975) in the High Arctic found that run-off was greater than recorded precipitation, and suggested that this problem was caused primarily by measurements of snowfall. However, due to problems in determining basin evaporation, it was not possible to determine the "true" snowfall as a residual of the water balance.

Although the processes responsible for this undercatch have long been known, the magnitude has been more elusive. Benson (1982) estimated that in northern Alaska actual snowfall was 3 times that reported by the U.S. Weather Service, and Adams *et al.* (1966) estimated that actual snowfall was 1.2 times greater than that measured in the Schefferville area. Woo *et al.* (1983a) conducted detailed snow surveys and found that snowfall in the High Arctic Islands was approximately 2 times higher than reported by the Atmospheric Environment Service weather stations at Resolute, Mould Bay, and Eureka, but that it varied from 1.3 to 3 times the measured amounts. These estimates have since been confirmed by detailed water balance studies (Woo *et al.*, 1983b).

Goodison and Louie (1985) discussed methods to correct the reported snowfall for stations in Canada, based on correcting gauge catch errors due to wind speed and the number of trace events. However, there have been no comprehensive studies to apply these corrections to standard weather station data in order to provide the best estimate of "true" snowfall in northern Canada. All of the published data and the snowfall maps (i.e., Figure 1.7) are based on uncorrected snowfall data or apply a single correction to all data (Maxwell, 1980). Unfortunately, the general attitude is "user beware".

In addition to the problems with underestimating point snowfall measurements in the Arctic, there is the additional problem with spatial variations due to the small number of stations. In the Queen Elizabeth Islands, for example, there are only 5 weather stations; all of which are at coastal locations. Given the small data base, snowfall maps such as the one shown in Figure 1.7, are typically constructed from the weather station data and then a simple relationship between elevation and precipitation is assumed in order to estimate snowfall in the mountainous areas. A classic example of the problems inherent in this method is the distribution of snowfall around the northern end of Baffin Bay. Koerner (1979) conducted an extensive survey of snowfall amounts on Devon, Ellesmere, and Axel Heiberg Islands and produced a map of snowfall over a 10-year period. This map is significantly different in both amount and distribution from the standard snowfall maps (i.e., Figure 1.7). For example, standard maps show peak annual precipitation of 200 $\text{mm}\cdot\text{yr}^{-1}$ coincident with the highest elevations. Koerner, however, found precipitation as high as 500 $\text{mm}\cdot\text{yr}^{-1}$ at low elevations in south-eastern Ellesmere Island and eastern Devon Island, with precipitation decreasing with increasing elevation and distance from the Baffin Bay polynya.

2.2.2. Snowcover Properties

Some of the earliest snow research in Northern Canada was aimed at the determining the physical properties of snow on the ground. For example, Williams (1957, 1958) and Longley (1960) provided the first description of snowcover properties such as density, grain size, and hardness; with particular emphasis on the changes over the winter and variation in properties from site-to-site. Further work in the 1960s and 1970s has provided more detailed descriptions of the regional variations in snow properties. For example, Pruitt (1970), who referred to the two main geographical snow types as tundra snow and taiga snow.

Tundra snow is dominated by blowing snow and thus is wind packed, has a high density, has a surface covered with dunes and sastrugi, and its distribution is highly variable. In fact the tundra snowcover is similar in many ways to the snow on polar ice sheets. Bilello's (1969) classification of snow density in North America, found that the mean seasonal snow density for the tundra regions (his category 4) was greater than 310 $\text{kg}\cdot\text{m}^{-3}$. Benson (1982) has shown examples of deep wind drift in river valleys on the Alaskan North Slope, while Woo *et al.* (1983a) showed that variations in snow depth and density were related to terrain type in the Canadian Arctic Islands. In this area, the snowcover was very thin (a few centimetres) on exposed hilltops, and extremely thick (many metres) in river valley and hillsides. These very thick valley and hill slope snowcovers may develop into perennial snowbanks, with an important effect on the run-off regime (Marsh and Woo, 1981), the geomorphic evolution of the terrain (Ballantyne, 1978), and may be important during the initial phases of glaciation. In an extensive study of tundra snowcovers in Alaska, Benson (1982) described four "types of snow" found in a typical profile. These include:

(1) fresh low density (150-200 $\text{kg}\cdot\text{m}^{-3}$) snow at the surface,
(2) fine grained (0.5 to 1 mm) dense (350-450 $\text{kg}\cdot\text{m}^{-3}$) wind slab,
(3) medium grained (1-2 mm), lower density (200-350 $\text{kg}\cdot\text{m}^{-3}$) snow, and
(4) depth hoar, with large grains (5-10 mm) and low density (150-300 $\text{kg}\cdot\text{m}^{-3}$).

In these tundra areas, the underlying soil is often very cold, with temperatures of -10 to -20°C or even lower. As a result the flux of heat and moisture from the soil to the snow during winter is generally small (Woo, 1982).

The second main type of snowcover is the taiga snow found in the forested areas south of the tree line. Blowing snow is not as important in the development of these snowcovers due to the lower wind speeds typical of the boreal-forest region and lower wind speeds below the forest canopy. As a result, these snowcovers are more evenly distributed than tundra snow, and are loosely consolidated, with a low density. Bilello (1969), for example, found that mean seasonal snow densities were less that 310 $kg{\cdot}m^{-3}$ in the northern forest of North America. The classic example of taiga snow is found in the interior of Alaska, where steep temperature inversions result in calm but very cold weather. Since the underlying soil is relatively warm, with typical temperatures of -5 to -10°C, and the snow surface often has very low temperatures (often below -50°C), extreme temperature gradients occur within the snowcover. Benson (1982) for example, described gradients of between 0.1 and 1.0$^{\circ}C{\cdot}cm^{-1}$. This results in large vapour-pressure gradients in the snow, and when combined with large heat and moisture fluxes from the soil (Santeford, 1979), the development of thick depth-hoar layers. In fact, the entire pack is often composed only of depth-hoar crystals (Benson *et al*., 1974).

In many boreal-forest areas, the taiga snowcover is not as well developed as that found in Alaska, and often is a mix of taiga and tundra snow. For example, in northern Quebec, the terrain is rolling, with boreal forests in the valleys and tundra on the hilltops. In this situation, the snowcover is extremely variable, with snow depth and density related to vegetation type (Granberg, 1988).

2.2.3. Snowmelt Run-Off

As the basic understanding of snowfall amounts and snowcover properties was improving during the 1950s and 1960s, researchers were turning their attention to the processes controlling snowmelt and snowmelt run-off. Energy-balance studies were conducted over both tundra (Weller and Holmgren, 1974; Ward and Orvig, 1953; Holmgren, 1971c; Price *et al*., 1976; and Heron and Woo, 1978) and taiga snow (Price *et al*., 1976). Weller and Holmgren (1974) and Heron and Woo (1978) showed that for tundra sites net radiation supplies 60 to 80% of the total energy available, with approximately 80% of the energy being used for melting the snow and the remaining 20% for evaporation. Heron and Woo (1978) found that energy-balance techniques accurately predicted melt when the tundra had a continuous snowcover, but that when the snowcover became patchy, the energy-balance approach became inaccurate due to the advective flux of sensible heat from nearby bare ground.

Price *et al*. (1976) provided one of the few detailed studies of melt rates for both forested and tundra sites in northern Canada. They found that winds at the tundra sites were approximately twice those in the forested sites, resulting in an approximate doubling of the sensible-heat exchange. For example, on a high-melt day the tundra site experienced 3.04 cm of melt due to sensible-heat flux, while the forest site had only 0.94 cm of sensible-heat melt. For net radiation, however, the opposite occurred, with net radiation being higher at the forest site than at the tundra site. For example, on the same day as discussed above, net-radiation melt was 1.05 cm on the tundra and 1.50 cm at the forest site.

The movement of meltwater from the snow surface to a stream channel, in both forested and tundra basins, has also been considered by a number of researchers. Price *et al.* (1976) conducted a detailed study of run-off from snow-covered, subarctic hillslopes which were underlain by permafrost. They provided one of the first applications of Equations 2.8, 2.9, and 2.10 to route water down a snow-covered hillslope. This modelling attempt was generally successful, but there were significant errors in estimating both the timing and magnitude of run-off. Price *et al.* (1976) described a number of the reasons for these errors, most notable being those dealing with the inability to handle the spatial variations in the snowcover, energy balance, and ground conditions.

Since the vertical flux of water through the snow can dominate the run-off process, Marsh and Woo (1985) conducted a detailed study of this vertical movement of water. They found that the flux of water is highly variable and that the snowcover cannot be considered homogeneous. This variability of flow is due primarily to the existence of ice layers which redistribute flow, concentrate it in some areas and lower it in others. The overall result is that water is able to move more quickly through the snow than is implied by Equation 2.8 and 2.9 for homogeneous snow. This process of flow through heterogeneous snow explains some of the problems encountered by Price *et al.* (1976) in applying Equations 2.8 and 2.9 to run-off forecasting. Marsh and Woo (1985) used a multiple flow path model which was able to accurately predict flow at the base of the snowcover.

A little-studied aspect of northern run-off processes is the initial movement of meltwater into dry, cold snow. Marsh and Woo (1984a, 1984b) found that Equation 2.2 can not be considered alone, but instead must be incorporated into a simulation model which links the movement of the wetting front to both the conduction of heat through the snow and soil, the development of flow fingers, and the growth of ice layers in the snow. Using a simulation model which linked a two-component wetting front, ice and basal ice growth, heat conduction through a connected snow/soil system, and soil infiltration, they were able to accurately predict the initial wetting of the snowcover, and the timing of the initiation of run-off.

Although many other studies have shown that the soil heat-flux term is small in some northern environments (Price *et al.*, 1976), Marsh and Woo (1987) found that in the High Arctic it is very large in magnitude and in fact plays an important role in controlling the release of meltwater from the snowcover. During the early stages of melt, the soil heat-flux term may in fact be larger than the surface melt-rate (Marsh and Woo, 1987). In this situation, if the soil infiltration is sufficiently small, meltwater ponded in a saturated layer at the base of the snowcover re-freezes to form basal ice. Further studies by Marsh (1988) have shown that in some circumstances the underlying soils have a very large infiltration capacity and as a result, basal ice does not form.

As suggested from the previous discussion, there have been numerous studies on many individual aspects of snowmelt run-off, but there have been few studies of the integrated snowmelt run-off system and issues such as the relative importance of various flow paths, source areas, or residence time of water in various storages. One of the most recent studies of this type was that by Obradovic and Sklash (1987) who used isotope and geochemical techniques to study the flow paths and residence times of snowmelt run-off waters in a 60 km catchment on Baffin Island. They found that 50% of the peak stream discharge was "old" water, where "old" water was defined as water added-to and stored in the basin during the previous summer, and "new" water is that from the melting snowpack. This result is rather unexpected for a permafrost basin where it is usually assumed that storage is minimal and that run-off from the snowpack dominates. This study clearly indicates that our present concepts of melt run-off are poorly tested in northern regions.

2.3. CURRENT RESEARCH

The following is a brief list of current snow-hydrology research as outlined by Winsor (1990) and Ommanney (1989), and known by the author to be ongoing in the Canadian North:

2.3.1. Blowing Snow

J. Pomeroy of the National Hydrology Research Institute (NHRI) has recently initiated a study on the physical processes of blowing snow in an arctic tundra environment and its effect on snow chemistry and snow properties. Field work is being conducted near Resolute Bay.

2.3.2. Meltwater Infiltration

P. Marsh (NHRI) and M.-K. Woo (McMaster University) have been studying the infiltration of meltwater into frozen soils at study sites near Resolute Bay. The principle aim of this study is to determine the role of infiltration in limiting snowmelt run-off.

2.3.3. Water Flux Through Cold Snowcovers

T.H. Illangasekare (University of Colorado) and colleagues have developed and are continuing testing of a 2-dimensional model of water movement in cold-glacier snow covers. Field work is being carried out on Ellesmere Island. The primary goal of this project is to obtain better estimates of the effect of global warming on run-off from polar glaciers, and the resulting lag time for sea-level rise. P. Marsh (NHRI), M.-K. Woo (McMaster University) and P. Barry (Atomic Energy of Canada Ltd.) have been studying the movement of wetting fronts into cold snow and the growth of ice lenses, ice columns, and basal ice in a cold snowcover near Resolute Bay. In addition they have been conducting tracer experiments to help determine the relative importance of pressure gradients and flow fingers in modifying the rising limb of meltwater hydrographs in melting snow. This project is aimed at improving techniques for modelling snowmelt run-off.

2.3.4. Remote Sensing

E. LeDrew (University of Waterloo) and colleagues have recently initiated a remote sensing study of snowcovers in the Arctic Islands near Resolute. A number of other studies, primarily linked through Canadian proposals associated with the Earth Observing System (EOS) satellites, are in a tentative stage of development.

2.3.5. Snow Chemistry

P. Marsh and M. Ferguson (NHRI) have been studying the movement of solutes through cold snowcovers in the High Arctic. This work was initiated to determine if the spring flushing of solutes observed in temperate snowcovers also occurs in cold snowcovers which experience considerable refreezing of meltwater during the melt period. D. Gregor (Water Quality Branch, Environment Canada) has been looking at the deposition of atmospherically-transported organic compounds in Arctic snow. Samples have been collected throughout the N.W.T. to obtain an estimate of spatial variation and from the Agassiz Ice Cap on Ellesmere Island in order to determine temporal variations over the last 17 years.

2.3.6. Snow/Ground Interface Temperature

H. Granberg (University of Sherbrooke) and D.T. Desrochers (McGill University) having been monitoring temperatures at a number of sites near Schefferville in order to provide data for use in spatial modelling of snowcover development and permafrost distribution.

2.3.7. Snow Radiative Properties

H. Granberg (University of Sherbrooke), A. Nadeau (McGill University), and A.V. Kulkarni (McGill University) have been conducting a detailed study of net radiation over snow at a subarctic forest edge in order to test models of net radiation within the forest. The attenuation of solar radiation by snow has been studied in order to develop a better understanding of the thermal regime of the seasonal snowcover.

2.4. FUTURE RESEARCH NEEDS

Although our understanding of snow hydrology has improved dramatically over the last 20 years there are still major gaps in knowledge that limit our predictive ability. Over the next 10 to 20 years there is going to be an increasing need to predict certain aspects of the snow environment. For example:

1. Water-resource studies will increasingly need to predict snowmelt run-off for water supply and hydro-electric development,
2. With the realization that arctic snowcovers are not pristine, but contain a range of industrial pollutants, there will be an ongoing concern about the release of pollutants from snowcovers to the arctic ecosystem, and

3. Global-change questions which will require prediction of large-scale snowcover and snowmelt run-off (Section 1.6.2.).

In order to meet these needs, the following research is required:

2.4.1. Snowfall

Snowfall amounts are still not well known in the Canadian North. Isolated studies have documented the magnitude of the undercatch at specific sites, but little is known about the spatial variations in the magnitude of gauge undercatch. As discussed earlier, undercatch is due to a combination of snowfall measurement during periods with high wind speeds and an inability to measure trace precipitation events. Extensive work has considered the effect of wind on gauge catch, but little work has considered the problem of trace precipitation. Previous attempts to account for trace precipitation (i.e., Bradley and England, 1978) arbitrarily assumed a single value (0.127 mm) for each trace event. It is still not known what percentage of annual snowfall occurs as trace events, the regional variations in trace precipitation events, or the processes responsible for these trace events. In general the number of trace events reported increase in a northerly direction. For example, from 1979 to 1989 the percentage of January days which had trace snowfall declined southwards from 72% at Resolute, to 44% at Inuvik, to 38% at Yellowknife. One possible explanation for this pattern is that, given the low air temperatures during winter (the mean minimum temperature in January is -36°C at Resolute, -34°C at Inuvik, and -33°C at Yellowknife), a large portion of the trace events are due to homogeneous nucleation (R. Stewart, AES, personal communication). If this is the case, and if these trace events are responsible for a significant portion of the total snowfall received in parts of the north, then trace events and snowfall amounts may be sensitive to small changes in air temperature. It is interesting to note that Bradley and England (1978) reported that at Resolute 49% of January days had trace precipitation during the period 1947 to 1972. This is considerably lower than for the period 1979 to 1989 noted above. However, the reasons for this change are not known. Since global-circulation models are predicting a significant rise in winter temperatures in polar regions, there could be a drastic decrease in the number of trace events. To fully understand the processes controlling trace precipitation and determine the percentage of snowfall which occurs as trace events requires joint work between cloud physicists and snow hydrologists. With the resulting advances in our understanding of the true magnitude of snowfall in the Canadian North, there is a need to adjust past precipitation records to reflect our best estimate of "true" values.

There is also a drastic need to increase the density of the precipitation network through the installation of remote stations or weather radars. Although this will never fully address the problems of spatial variability, it will provide a better ground "truth" data base for comparison of snowfall/snowcover data which may become available after the launch of the EOS (Earth Orbiting System) satellites.

2.4.2. Snowcover

Improved understanding of the processes controlling, and models of, snowcover properties during the winter are required for calibrating remote sensing signals, for predicting ground temperatures and permafrost under climatic-warming scenarios, and as input to snowmelt run-off models. Among the most important requirements are:

1. To model metamorphic changes that occur within the snowcover during the winter. There are empirically-based models available (i.e., Anderson, 1976), but these have not been tested extensively in the North. There is a need for physically-based process models that would include changes in snow density, grain size and shape, development of layers within the snowcover, and the vapour and heat fluxes between the snow and underlying soil.
2. To model snowcover distribution within drainage basins. This is not a significant problem in forested areas, but is a major problem in tundra areas where the snowcover is highly variable due to blowing snow. We need to be able to predict snowcover distribution for different terrain and climatic conditions. This will require an improved understanding of blowing-snow processes, the development of large-scale flow models for predicting wind over rough terrain, and the linking of metamorphic and blowing snow models.
3. An improved understanding of continental-scale variations in snow properties, including snowcover depth, density, water equivalent, grain size, stratification, liquid-water content, and the development of physically-based parameterization techniques for use in continental-scale models. Complete success will require the development of appropriate remote-sensing and modelling techniques.

2.4.3. Snowmelt

Although energy-balance techniques are commonly used to predict snowmelt rates at a point, there has been little progress made on extending these techniques to the regional scale. Radiation measurements over large areas can be obtained by a combination of remote sensing and modelling. Estimates of sensible- and latent-heat fluxes are more difficult to obtain over rough terrain. This problem can only be solved by improvements in modelling techniques. A further problem in using energy-balance techniques in tundra areas is the problem with advection to residual snowbanks.

2.4.4. Snowmelt Run-Off

There is a need for studies of snowmelt run-off from larger drainage basins in order to determine the importance of various flow paths and source areas, and residence times. These types of studies can only be successful through an integrated approach using physically-based models, isotopic and geochemical tracers, and remote sensing in conjunction with detailed ground-truth data.

Although there are numerous snowmelt run-off models available, none were especially designed for use in northern regions. There is a need for extensive testing and possible modification of existing models and/or development of a new physically-based, snowmelt run-off model designed to handle the unique aspects of northern hydrological systems. These include:

(1) large negative soil heat flux,

(2) infiltration into frozen soils,

(3) infiltration of meltwater into cold snowcovers, and

(4) development of ice layers within and at the base of the snowcover.

In addition to these features, there is a need to improve our understanding of a number of basic processes and relationships that are not unique to the North but are common to all snowmelt run-off systems. These deficiencies must be addressed before comprehensive physically-based snowmelt run-off models can be developed, and before appropriate parameterization techniques can be incorporated into large-scale models (either hydrological or global-climate models). For example:

(1) the relationship between grain growth rates and water saturation,

(2) the hydraulic properties of wet snow, including the relationships between non-saturated permeability, water saturation, and water pressure, and

(3) the processes responsible for the development of flow instabilities at the leading edge of wetting fronts.

2.4.5. Removal of Solutes from Melting Snowcovers

With increasing northern development and concern for long-range transport of pollutants, there is a growing realization that snowcovers in northern Canada contain a range of industrial pollutants. There is therefore a developing need to understand the pathways that these pollutants take from the snowcover to streams and lakes. For temperate snowcovers, it is well known that there can be an initial pulse of meltwater with a concentration of pollutants. It is not known how this phenomena is modified in cold snowcovers where there is considerable freezing of meltwater within and at the base of the snowcover, and in the underlying soil. *

3

PERMAFROST HYDROLOGY

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3.1. INTRODUCTION

Permafrost hydrology studies the distribution, movement and storage of water as it is directly or indirectly influenced by the presence of perennially frozen ground (Woo, 1986; permafrost is defined in Section 1.3.2.4.). The formation and maintenance of permafrost is due to intense and persistent cold conditions, and therefore the prevalence of coldness permeates the hydrological cycle in the North. Chapter 4, for example, provides a detailed review of the role of permafrost in the northern ground-water regime.

Permafrost underlies over half of Canada (Brown, 1970) and embraces a range of geology, topography and vegetation (Figure 1.11; Section 1.3.2.4.). Broadly speaking, the zone that is underlain by continuous permafrost retains a snow and ice cover for at least eight months of each year (Canada, Fisheries and Environment, 1978). The southern limit corresponds roughly with the tree line, but based on the vegetation cover, the continuous permafrost region can be subdivided into three zones (Bird, 1967). The High Arctic has barren areas with patchy tundra vegetation; the Middle Arctic has tundra mixed with bare ground, and the Low Arctic has a continuous tundra cover. The discontinuous permafrost zone is largely covered by boreal forests mixed with extensive wetlands (Bliss, 1979; National Wetlands Working Group, 1988). The number of snow-free days averages six months at the southern fringe where permafrost occurs sporadically.

Despite the diversity of its physical environment, the permafrost region shares several common traits in its hydrological behaviour. (1) The frozen ground has limited permeability and it acts effectively as an aquiclude (Dingman, 1975)(aquitard, see Section 4.1.2.2..). (2) Most hydrological activities are confined to the seasonally-frozen and thawed zone above the permafrost table, known as the active layer. (3) Many surface hydrological processes are inactive during the long, cold winters. (4) Energy and water fluxes are closely linked as water storage and redistribution are modified by freeze-thaw events. (5) Snow and ice storage on a seasonal or multi-annual basis affects the temporal distribution of water, and the release of meltwater often has pronounced effects on other surface hydrological processes. This chapter deals with the surface component of permafrost hydrology, but it is clear that through the movement and storage of water, the surface component is coupled with the ground water (Chapter 4), snow

(Chapter 2) and atmospheric aspects (Chapter 8) of the hydrological cycle. In this regard, there are strong linkages between this and other chapters of this book.

Hydrological research in the North is driven by scientific curiosity as well as by the pressure for applications. The pronounced spring freshet in permafrost areas is a frequent cause for flooding and severe erosion. The high degree of saturation of the active layer during the melt season renders the slopes extremely vulnerable to irreparable damage, but the long frozen season permits winter roads to traverse terrain that is inaccessible to vehicular traffic during the thawed period. Many northern wetlands exist because of an impervious frozen substrate and these lands of abundant surface water offer vital habitats for many wildlife and waterfowl. On both a regional and a local basis, an inventory of the water resources is basic to solving the water-supply problems of northern communities, and for the long-term planning for regional development. As human activities increase in the North, the hazard of pollution increases and permafrost hydrology offers an important key towards preserving the environment for the generations to come.

3.2. HISTORICAL BACKGROUND

Research in permafrost hydrology has been restricted by (1) difficult or expensive access to the sparsely-populated permafrost areas, and (2) the low demand for extensive development in the region, and hence inadequate resources to carry out research. A lot of the work is piecemeal in nature, or has a short life span. The lack of long-term commitment to permafrost hydrological research, mainly in terms of funding, is aggravated by an inadequate hydrometric network and the scattered climatological stations sited in northern Canada (Section 1.3.2.8.). In spite of these difficulties, notable advances have been made in the last two decades.

Many lines of permafrost hydrological enquiry have evolved through the stages of phenomenon identification, measurement and empirical correlation, followed by process investigation. Sometimes quantitative observations led directly to predictive procedures when demands for immediate application pre-empted the thoroughness of scientific endeavour. The history of hydrological investigations in the North is a mixture of opportunistic observations by non-hydrologists, the rise and fall of engineering concerns as development proposals waxed and waned, and scientific research which followed the special interests of individuals or their groups.

The ground-water aspect received early attention because of its practical application and the impetus from the Soviet Union and the United States (Tolstikhin and Tolstikhin, 1976; Williams, 1965). It is to be noted that ground water was the primary topic in permafrost hydrology to appear in the early Proceedings of the International Conferences on Permafrost (U.S. National Academy of Sciences, 1966; 1973). This aspect is reported in Chapter 4 and will not be elaborated on here. To trace the developmental history of permafrost hydrology in Canada, one can refer to the series of review articles by Church (1974), MacKay and Løken (1974), Anderson (1980), Woo (1986), and van Everdingen (1987).

3.2.1. Streamflow Studies

Streamflow is an obvious hydrological phenomenon to be examined. Cook (1967) measured the flow of Mecham River in the continuous permafrost area of Cornwallis Island, and this was followed by more a detailed comparison of diurnal streamflow cycles with the radiation (snowmelt) patterns (McCann and Cogley, 1972). Church (1972) carried out similar observations in proglacial areas of Baffin Island. A spate of streamflow measurements followed in the early 1970s (e.g. Ambler, 1974; Anderson, 1974; Holecek and Vosahlo, 1975). The 1970s also witnessed an expansion of the hydrometric network in the North (Canada, Environment Canada, 1988b).

The role of permafrost in basin hydrology is to accentuate run-off peaks while reducing base-flow contribution (Slaughter *et al.*, 1983). By 1974, Newbury (1974) concluded that permafrost basins behave like impermeable catchments in terms of their rainfall/run-off relationship. Within the extensive permafrost region, variations in run-off patterns exist because of differences in the local environment. Based on data from Canada and Alaska, it was possible to generalize streamflow regimes in permafrost areas into the four categories described in Section 1.4.1. and illustrated in Figure 3.1. Because of the importance to this discussion, they are reviewed below in more detail.

Nival regimes are dominated by snowmelt floods in spring, and can be distinguished into a subarctic and an arctic category. The former experiences limited winter low flow sustained by the discharge of ground water, and a spring freshet related to ice jams (Section 5.3.1.). Summer usually has low flows but large peaks may be generated by heavy frontal rainfall in the discontinuous-permafrost region. The arctic-nival regime shows major floods caused by snowmelt, and usually accentuated by the presence of snow jams in the channels (Woo and Sauriol, 1980; Figure 3.2). Spring high flows are followed by rapid recession to base flow (Figure 3.2), which is occasionally punctuated by rainstorm-generated peaks. Winter flow is absent because the suprapermafrost can alter the regime considerably (Figure 3.2). The normal expectation is that lakes would regulate run-off by reducing peak flow and maintaining low flow. Where the lake outlets are blocked by snow at the end of winter, the snow dams can impound much meltwater on the lake ice before being breached to release an extraordinarily large flood.

Glacier meltwater-fed rivers exhibit a proglacial streamflow regime with spring snowmelt run-off superseded by summer peak flows from glacier melt (Figure 3.1c). As noted earlier (Section 1.4.1.3) nival regimes are not always easy to distinguish from the proglacial regime because snowbanks at higher elevations may maintain high flow during summer in a manner analogous to glaciers (Marsh and Woo, 1981). In some basins, polar glaciers impound ice-dammed lakes that burst at irregular intervals to generate exceptionally large floods called jøkulhlaups (Blachut and McCann, 1981; Maag, 1969; Sections 6.2.5. and 6.5.3.2.).

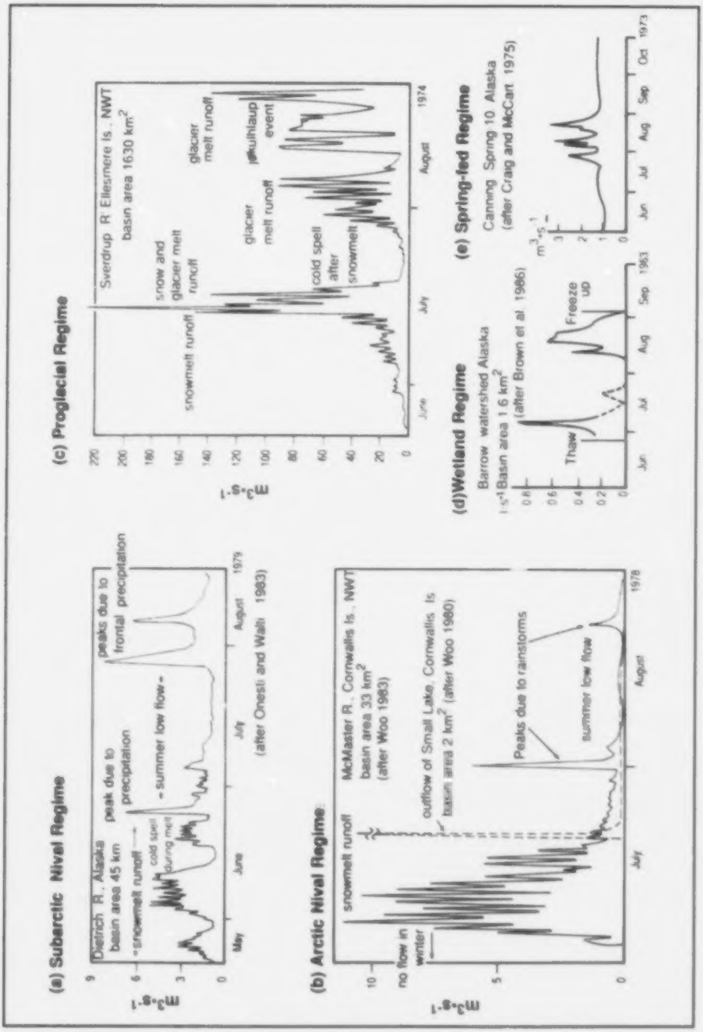

Figure 3.1. Hydrographs representing several typical streamflow regimes in permafrost areas:
(a) subarctic-nival regime
(b) arctic-nival regime, also shown is outflow from an arctic lake
(c) proglacial regime
(d) wetland regime
(e) spring-fed regime *(Woo, 1986).*

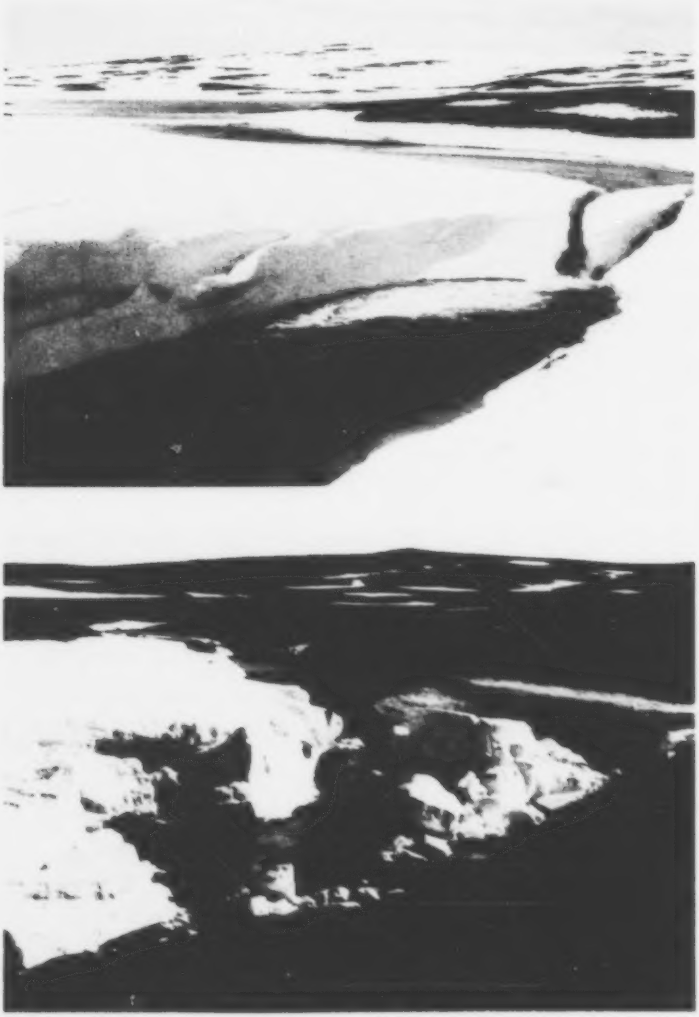

Figure 3.2. Flow of McMaster River showing typical arctic-nival regime run-off.
(a) snow dam dissected by two channels during the snowmelt period (July 7, 1978), draining the temporary snow-dammed lake behind and causing peak flow downstream
(b) mid-July low flow; photograph was taken at the same position and showing decay of the snow dam (↓); shading shows the location of the drained snow-dammed lake.

The wetland regime, called a muskeg regime by Church (1974), includes the regime of tundra streams of Craig and McCart (1975). The run-off pattern is governed by prolonged seasonal frost, rapid and considerable release of snowmelt water, limited ground-water storage capacity and the prevalence of surface flow on wetlands (Woo, 1988). When the ground is frozen during the snowmelt period, surface flow sweeps rapidly across the wetlands as soon as depression storage is filled. Afterwards, when the ground thaws and as the water recedes, greater water-retention capacity becomes effective and summer floods are less intense than those experienced by nival-regime rivers.

In addition to the four basic classifications noted by Church (1974), Craig and McCart (1975) also recognized a spring-fed regime (see Sections 4.3.4.1. and 4.3.5.4.) Often associated with carbonate rocks in the discontinuous-permafrost zone, such streams have relatively stable discharges because the primary source of water supply is from ground water. Snowmelt and intense rainfall can cause flooding to many karst depressions as is shown by the hydrograph of Snag Creek (Figure 4.9) while continuous winter discharge can lead to icing formation downstream from the springs.

3.2.2. Water-Balance Studies

Closely associated with streamflow studies were such observations as precipitation, snowcover and evaporation, couched in a basin water-balance framework:

$$P = R + E + \Delta S \qquad [3.1]$$

where P is precipitation, R is run-off, E is evaporation and ΔS is change in storage. An early review of North American contributions in this regard was given by Dingman (1973a). Up to that time, no study dealt with all components of the water balance, always leaving at least one term as the residual of computation. This means that the errors attributed to the measurements of the known terms cannot be separated from the magnitude of the residual term.

Water-balance studies were made at different parts of the permafrost region in the 1970s (Cogley and McCann, 1975; Holecek and Vosahlo, 1975; Newbury *et al.*, 1979; Wedel *et al.*, 1977b; Wright, 1981). To be precise in the water-balance measurements, however, it is necessary to study the temporal and spatial variations of the various hydrological elements and to understand how the hydrological processes operate in the permafrost environment. Process studies at a site-specific level increased during the 1970s. The impetus came partly from northern oil and gas development; partly as a result of inter-disciplinary programmes such as the International Biological Programme; and partly through the research interest of several institutions. Notable amongst the processes studied were snowmelt (Heron and Woo, 1978; Moore, 1983; Price and Dunne, 1976), evaporation (Addison, 1977; Marsh *et al.*, 1981; Ohmura, 1982b; Rouse *et al.*, 1977), and soil moisture redistribution during freeze and thaw (Jame and Norum, 1980; Sheppard *et al.*, 1978). The soil-moisture aspect is described in Chapter 4 and will not be discussed in the surface-water context.

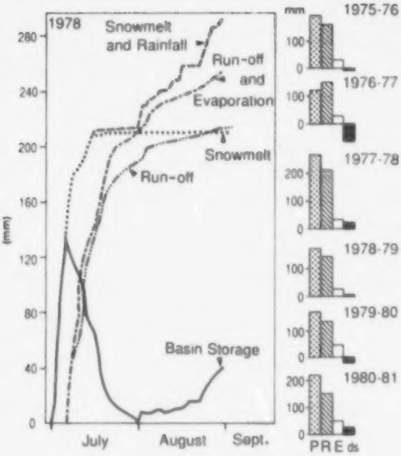

Figure 3.3. Daily water balance for McMaster River basin during the summer of 1978; and annual water balance for 1975-81. (P = precipitation, R = run-off, E = evaporation, ΔS = change in storage).

On a basin level, the spatial variability of various processes was examined. An example was given by Woo *et al.* (1983b) in their study of the water balance of McMaster Basin (area 33 km^2) between 1975-81. The basin was divided into several terrain units; snow surveys were carried out at the end of each winter to determine the total accumulation and distribution of snow in various types of terrain. During the spring, the changing combination of snow distribution and evaporative snow-free surfaces was used to investigate the basin gains and losses (Woo and Heron, 1979): melt quantity was weighted by the snow distribution on a given day, and evaporation was computed for those terrain units that became snow-free. Summer rainfall was measured by a network of gauges and streamflow was obtained at the outlet of the basin. Water balance was then computed on a daily basis, and a 'water year' was considered to span the time between the first significant snowfall in September of the first year and the arrival of snowfall a year later. Figure 3.3 presents the cumulative daily snowmelt, rainfall, evaporation, run-off and change in storage (computed as a residual) for the 1977-78 'water year'. This example shows that snowmelt began late in 1978 (July 1), but three-quarters of total winter snow accumulation was lost within one week. Evaporation was negligible until run-off began on July 6 when an integrated-channel flow network was established in the snow-filled valleys (Woo and Sauriol, 1980) to drain most of the water stored in the basin. Evaporation, though low in magnitude, increased gradually. Storage was much reduced by late July and dry conditions prevailed. Rainfall in August occasionally raised the streamflow, but a sizeable portion of the precipitation went into replenishing the basin storage. Streamflow and evaporation declined by September, as freeze-up began. An excess of storage at the end of summer extended into the subsequent spring, showing that the net change in storage at the end of the year was not negligible.

Water-balance studies in non-glacierized, permafrost basins yield the following generalizations.

(1) Snowfall constitutes the bulk of annual precipitation and it is of utmost importance to determine the total basin snow storage. Weather-station snowfall data notoriously under-estimate the basin snow accumulation and have led to many earlier problems in water-balance research.

(2) Most of the annual run-off in the Arctic is fed by snowmelt (hence, the term nival regime), but the importance of snow declines southward where rainfall contributes larger portions of the precipitation: then the pluvial run-off regime dominates.

(3) Run-off consumes a large portion of precipitation. Typical values of the run-off ratio (R/P) range from 0.7 to 0.8 for non-wetland basins (Anderson, 1974; Findlay, 1969a; Woo, 1983), but in basins with much wetland or open-water surfaces, the evaporation rate becomes relatively high and the ratio drops to about 0.5 (Rydén, 1977).

(4) Although the active layer offers limited storage capacity, the year-to-year change in basin storage can account for 10 percent of the annual precipitation (Figure 3.3). It is therefore erroneous to assume, as did the earlier water-balance works, that the basin-storage change from year-to-year is minimal.

3.2.3. Hillslope Hydrological Studies

Basin research was overlapped by hydrological investigations on the scale of hillslopes. In permafrost areas, the proximity of an impermeable, frozen substrate near the ground surface often reduces the storage capacity of the active layer, and lateral water movement is enhanced. The processes leading to run-off generation and the movement of water within the active layer or on the slopes have been studied in the Subarctic (Wright, 1979), Middle Arctic (Lewkowicz and French, 1982) and in the High Arctic (Woo and Steer, 1982, 1983).

Water flow on slopes commonly commences during the snowmelt period. Initially, run-off occurs as flow through snow. The presence of a frozen substrate close to the ground surface hinders percolation at a time when water supply is abundant (Wilkinson and Bunting, 1975; Woo, 1976b) and soon the depression storage requirements are satisfied, followed by the occurrence of overland flow (Figure 3.4). Water moves as sheets or rills on the slope surface. Subsurface flow increases gradually as the active layer thaws to allow greater vertical and lateral water movement (because hydraulic conductivity increases by several orders of magnitude when the frozen material thaws, see Burt and Williams, 1976; or Kane and Stein, 1983b). On different segments of a slope, surface and subsurface flows can occur together or they can alternate, depending on whether water emerges or submerges in the active layer. The mechanism of flow production was conceptualized by Dingman (1973b) that, during the wet period, a rise of the water-table above the slope in low-lying areas causes surface run-off, while in the dry period, the shrinkage of overland flow areas corresponds to a lowering of the water-table below the surface of the ground.

Figure 3.4. Extensive overland flow on hill slopes of central Ellesmere Island during the snowmelt period.

In the Subarctic, a contrast in slope aspects can cause a systematic variation in permafrost distribution, hence local differences in slope hydrology. South-facing slopes have a deep active layer or have no permafrost and water can enter the mineral soil without yielding significant lateral flow. But on north-facing slopes, deep percolation is restricted by the essentially impervious frozen substrate, and lateral flow dominates to give rapid and significant run-off responses to snowmelt and to rainfall (Slaughter and Kane, 1979). In continuous permafrost areas, permafrost underlies all slopes at shallow depths, but differences in the radiation balance may produce significant differences in snowmelt and evaporation regimes. The water-table is often closer to the surface of north-facing slopes, and the near-surface moisture content can be lower for the south-facing slopes (Flügel, 1983). Regardless of slope direction, concave slope segments often have a higher water-table and trap more snow. These are areas favourable to the maintenance of late-lying or semi-permanent snowbanks which continue to supply water to slope run-off long after the spring season (Young and Lewkowicz, 1988).

3.2.4. Studies of Lake and Wetland Hydrology

Lakes are numerous in the permafrost environment. Limnologists have always had peripheral interests in the hydrology of lakes and their basins. The Char Lake project, carried out near Resolute in the early 1970s, provided information on arctic lake characteristics (Schindler *et al.*, 1974b). Hydrological studies of northern lakes include lake evaporation (Louie, 1979; Marsh and Bigras, 1988; Roulet and Woo, 1986b; Stewart and Rouse, 1976b), lake level fluctuations (Marsh, 1986), lake discharge (FitzGibbon and Dunne, 1981; Woo, 1980), and lake ice (see Chapter 5). Several of the studies also include the water balance of the lake or pond and its drainage basin (Marsh, 1986; Marsh and Woo, 1977; Woo *et al.*, 1981).

The North abounds in wetlands (Section 1.3.2.3.), many of which are underlain by continuous or discontinuous permafrost (National Wetlands Working Group, 1988); but their hydrological behaviour is not well known (Ford and Bedford, 1987). Based on the limited amount of information available (Roulet and Woo 1986a; Roulet and Woo, 1988; Rydén, 1977; Woo, 1983), several generalities can be surmised.

(1) The frozen condition of wetlands during spring permits little snow meltwater to infiltrate. Surface flooding recurs annually as does melt-related peak flow in wetland streams.

(2) Moisture is generally available near the surface and summer evaporation is enhanced.

(3) The ability of the wetlands to regulate streamflow depends on the availability of storage capacity for water input (such as rainfall, snowmelt and lateral inflow).

When the water-table is close to the surface, there is little capacity to absorb additional water and run-off response is rapid; but prolonged evaporation and continuous thawing of the active layer increases the storage capacity so that high flows are attenuated and low flows are extended. Roulet and Woo (1988) provided an example of a basin in Keewatin that consists of lakes, hillslopes and wetlands. They showed that run-off production is controlled by the spatial integration of these distinct hydrological systems, and the linkages of these systems change from one period to another in the course of a summer.

3.3. RECENT TRENDS

By the mid-1980s there was a decline in permafrost hydrological research. The major physical processes were recognized and largely understood. The principles governing small-basin responses to snowmelt and rainstorms in the North were known in general terms. The lull in field research provided an opportunity to assess the achievements over the past two decades and direct attention towards fresh hydrological initiatives.

3.4. CURRENT RESEARCH

As indicated by the paucity of papers published on permafrost hydrology in the Proceedings of the 5th International Conference on Permafrost (Senneset, 1988), Canadian contributions to surface hydrology have dwindled. Current efforts include studies of several hydrological processes not previously completed, and the consolidation of existing knowledge into modelling structures.

3.4.1. Coupling Snowmelt, Infiltration and Soil Moisture

Processes controlling the movement of meltwater through a cold snowcover have been discussed in Chapter 2. When the meltwater reaches the base of the snowpack, infiltration into the frozen ground may occur (Gray *et al.*, 1986; Kane and Stein, 1983b). The quantity of meltwater infiltrated may be so large in certain types of soil that little surface run-off is produced (Marsh, 1988).

The soil-moisture regime during the winter and the spring (Smith, 1985) has important implications for melt run-off and for the thermal conditions of the active layer. The present lack of knowledge in this respect is one of the factors limiting the use of existing snowmelt/run-off models in northern basins. Field experiments are being conducted by the National Hydrology Research Institute and by McMaster University at a subarctic (Inuvik) and an arctic (Resolute) site to trace the pathway of meltwater from the snow to the ground. The emphasis is on infiltration into frozen ground as influenced by winter soil-moisture and thermal conditions, and by the delivery of meltwater in the spring. Gamma probes are used to determine moisture content both in the frozen and unfrozen states, while time-domain reflectometry is used to detect the unfrozen water content in the same soil profiles. Meltwater will be routed through the snowpack to the ground (Marsh and Woo, 1984a) to be coupled with the observed soil-moisture changes and the subsequent refreezing of the infiltrated water. Several types of soils are monitored to obtain the spatial variability.

3.4.2. Hydrological Modelling

Modelling had been applied to facets of permafrost hydrology such as the soil moisture and ground temperature regimes (Harlan, 1973; Outcalt *et al.*, 1975), but an attempt at comprehensive modelling of small-basin run-off met with little success (Ambler, 1979). The reasons were the lack of adequate input data and, more seriously, the use of a model developed for the temperate latitudes which did not include many of the processes relevant to the permafrost environment (e.g., a dynamic thawing zone in the active layer). Another approach was to treat the basin as a black box and predict design discharge from meteorological records using statistical methods (Ashton and Carlson, 1983; Baracos *et al.*, 1981).

Scientific hydrological studies have yielded sufficient information to allow an attempt at producing physically-based, deterministic models tailored to the permafrost region. Hydrometeorological data remain limited, however, because of the sparse weather-station network. To be practical, a model has to make use of readily-obtainable data including air temperature, precipitation and perhaps, radiation. A one-dimensional model is proposed by Woo and Drake (1988) to simulate the daily hydrological and thermal conditions of the vertical column (Figure 3.5). The snow on the ground is incremented when air temperature is below freezing on a day with precipitation. Snowmelt is partitioned into radiation melt, rain-on-snow melt and turbulent-flux melt indexed by air temperature (because wind and humidity data are generally unavailable). The penetration of the meltwater front utilizes

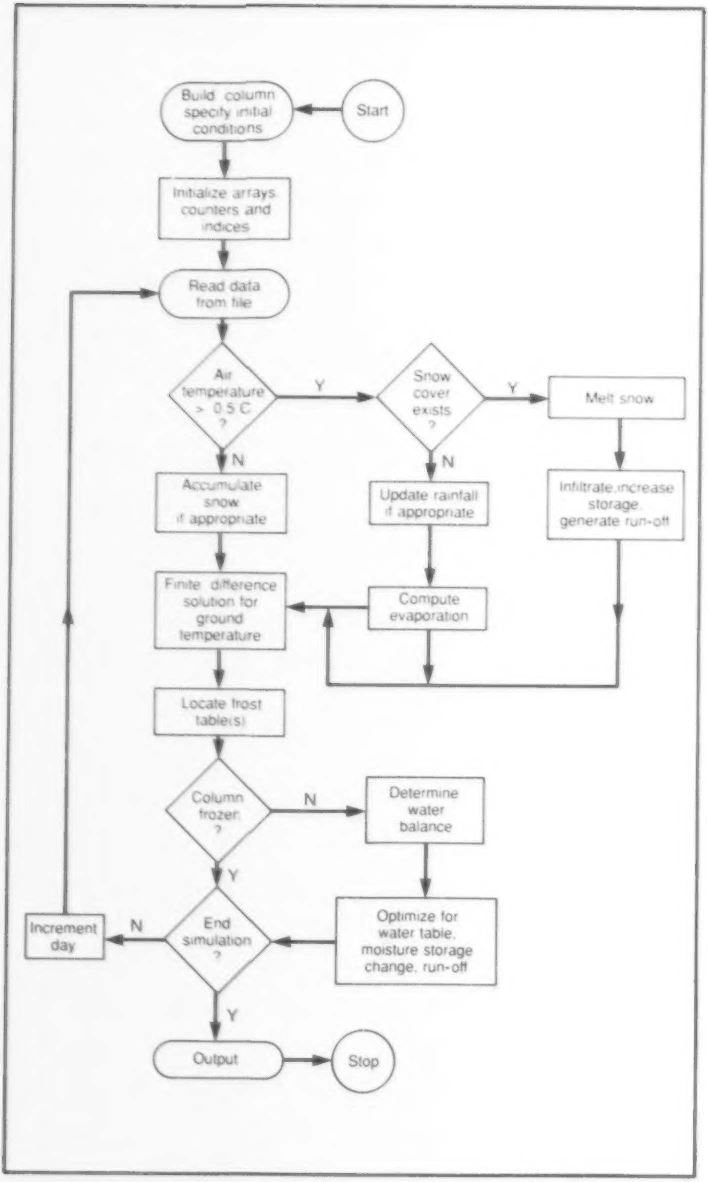
Start
Build column specify initial conditions
Initialize arrays counters and indices
Read data from file
Air temperature > 0.5 C ?
Y
Snow cover exists ?
Y
Melt snow
N
Accumulate snow if appropriate
N
Update rainfall if appropriate
Infiltrate, increase storage, generate run-off
Finite difference solution for ground temperature
Compute evaporation
Locate frost table(s)
Column frozen ?
N
Determine water balance
Y
Increment day
N
End simulation ?
Optimize for water table, moisture storage change, run-off
Y
Output
Stop

Figure 3.5. Flow chart for the modelling of hydrological and thermal processes at a permafrost site *(Woo and Drake, 1988)*.

Colbeck's (1976) equation and basal ice growth and infiltration into frozen soil are considered in a primitive fashion. Evaporation is computed using the Priestley and Taylor (1972) approach. Ground temperatures are calculated using finite difference, and the solution allows the frost-table to be located. Daily water balance is performed and the water-table is updated. When the water-table rises above the ground to exceed the depression storage, lateral run-off is generated from the site.

This algorithm can be used to generate run-off for different points in the basin or on a slope, but further work is needed to bring the point values into a spatial framework, and to route the flow down the slopes and along the channels. The finite-element approach may be used to bring a spatial dimension to the point model. Basin outflow may not be modelled easily during the break-up period because the channels are choked with snow and ice. Advances in river-ice research (see Chapter 5) will hopefully provide the information essential to model the channel phase of permafrost hydrology.

3.5. FUTURE RESEARCH NEEDS

Permafrost hydrological research can benefit from broadening the spatial dimensions to encompass regional characterization, and from expanding the scope temporally to predict possible changes in response to the future climate.

3.5.1. Regional Hydrology

Insufficient attention has been paid to quantitative characterization of permafrost hydrology on a regional scale. This is of considerable importance in terms of regional planning and engineering applications. The tool is statistics but the support is a broad hydrometric data base. The major problems have been the lack of long-term records, the unreliability of high-flow data, and a scanty gauging network. Yet, it is precisely because of these difficulties that statistical analyses of the available information should be performed to extend the records, to fill in the missing values and to make spatial interpolation to ungauged basins. Some work has been done previously to estimate floods for arctic gas pipeline river crossings (Solomon *et al.*, 1977). With a larger data base (Canada, Environment Canada, 1988b), the regional analysis of different aspects of the run-off regime can be expanded.

3.5.2. Climatic Change and Permafrost Hydrology

As the atmospheric concentration of carbon dioxide and several trace gases increase, global warming appears to be inevitable. Potential changes to the permafrost hydrological environment have been reviewed in Section 1.6.2.3. There is as yet no definite evidence that a warming trend is descending upon the North, but it is too risky not to heed these possibilities and be prepared for such contingencies.

Given our present knowledge of permafrost hydrology, we can conjecture various effects of climatic change. However, predictions should be made with some specified degree of confidence and further research is required. To do so, three approaches are recommended.

3.5.2.1. Application of Spatial Analogues

The present-day hydrological conditions at temperate latitudes, for example, may be used to predict what the Subarctic could be when the climate changes. Caution must be exercised because the solar-radiation regime will remain largely unchanged and vegetation feedback will lag behind climatic change so that spatial equivalence need not apply.

3.5.2.2. Application of Climatic Variability Analogues

Hydrological information for the warmer years can be used to infer what may happen as global warming occurs. Due consideration must be given to the unknown persistence effect when short-term climatic fluctuation observations are used to estimate the results of long-term climatic change.

3.5.2.3. Use of Simulation Models

By putting together various hydrological processes in a model and applying new climatic forcing functions, future hydrological conditions in permafrost regions can be simulated quantitatively. Care must be given to ensure that the models have a good physical basis and that the results are realistic.

These approaches are not mutually exclusive and all should be attempted to obtain a firm grasp of the climatic change effect. Findings from these studies will be of importance to the International Geosphere and Biosphere Program which has a global mandate. ❄

4

GROUND-WATER HYDROLOGY

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4.1. INTRODUCTION

4.1.1. Permafrost

The presence of permafrost influences the occurrence, movement, and quality of ground water to various degrees over about half the land area of Canada (see definitions and other descriptions in Sections 1.3.2.4/5). It occurs in areas with mean-annual air temperatures below -0.5°C, and is generally overlain by a layer of variable thickness, subject to seasonal freezing and thawing, known as the active layer. The relationship between the two, and their dependence on temperature variations with depth are shown in Figure 4.1. As described in Section 1.3.2.4, permafrost underlies much of the arctic and subarctic regions of northern Canada, and parts of the shallow offshore in the Canadian Arctic; it is also found in alpine zones on high mountains (Figure 1.11). Permafrost is discontinuous in the southern portions of the permafrost region; it may occur as thin isolated masses of perennially-frozen ground near the southern fringes of the discontinuous zone. Permafrost becomes gradually thicker and more continuous with increasing latitude, as a consequence of decreasing mean-annual air temperatures, but non-permafrost "windows" occur even in the zone of continuous permafrost. Local distribution of permafrost is controlled by elevation, exposure, vegetation, extent of snowcover, and presence of rivers, lakes, and glaciers (Brown, 1970).

4.1.2. Properties of Frozen Ground and Permafrost

Freezing of a water-bearing soil or rock affects its physical properties: electrical and hydraulic conductivity decrease by several orders of magnitude; apparent specific heat capacity decreases; and thermal conductivity increases (Anderson and Morgenstern, 1973). Any water remaining in the liquid phase (e.g., due to depression of its freezing point by dissolved-mineral content) will exhibit increased viscosity.

4.1.2.1. Ice and Water Contents

It is often assumed that frozen ground contains water only in the form of ice; that frozen ground is therefore impermeable; and that permafrost is perennially-frozen and, thus, impermeable. These assumptions are not always valid.

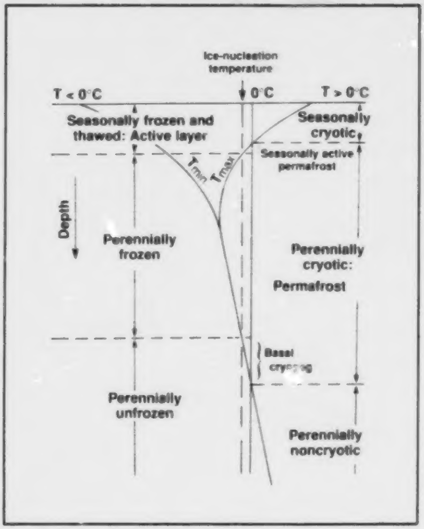

Ice-nucleation temperature
T < 0°C
0°C
T > 0°C
Seasonally frozen and thawed: Active layer
Seasonally cryotic
Seasonally active permafrost
T_{min}
T_{max}
Depth
Perennially frozen
Perennially cryotic:
Permafrost
Basal cryopeg
Perennially unfrozen
Perennially noncryotic

Figure 4.1.
Schematic representation of the variation of temperature with depth in permafrost, showing terms indicating the seasonal or perennial temperature relative to 0°C, and terms indicating seasonal or perennial presence or absence of ice *(van Everdingen, 1987).*

Liquid water exists in many soil-water-ice systems, as thin films adsorbed onto the surfaces of soil (mineral) particles, even when the temperature is well below the initial freezing point of the soil water (see Figure 4.2). The thickness of the water films is a function of temperature (50 Å or more at 0°C; 9 Å at -5°C; 3 Å at -193°C); it is nearly independent of the total water content (liquid plus solid), except in very dry soils (Anderson and Morgenstern, 1973). The adsorbed water is mobile under both electrical and thermal gradients. As unfrozen water content is primarily a function of temperature below 0°C, a thermal gradient in a frozen porous medium is analogous to a water-content gradient in an unfrozen, unsaturated porous medium.

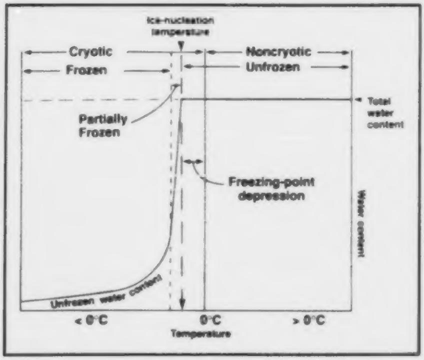


Figure 4.2.
Hypothetical graph of unfrozen water content vs. ground temperature, showing the relationship between terms indicating temperature relative to 0°C and those indicting phase conditions *(van Everdingen, 1987).*

Permafrost, defined solely on the basis of temperature, is not necessarily frozen. It may be "dry", containing neither water nor ice. It may contain unfrozen water exclusively, if the freezing-point depression is large enough to prevent freezing at the prevailing negative temperature; it may contain a mixture of unfrozen water and ice; or it may contain ice exclusively. Segregated ice, in the form of lenses, layers, or ice wedges, may also be present.

4.1.2.2. Hydraulic Conductivity

Frozen and unfrozen portions of the same unsaturated soil may have similar hydraulic conductivities when soil temperatures are close to 0°C (Anderson and Morgenstern, 1973). With decreasing temperature below 0°C, the available cross-sectional area of interconnected films of unfrozen water decreases, and so does the hydraulic conductivity of the soil.

In frozen soil materials, the rate of water movement thus depends on the overall temperature of the system, on the thermal gradient, and on the available cross-sectional area of interconnected films of unfrozen water. The latter is commonly largest in fine-grained materials (silts and clays).

Frozen ground, whether seasonal or perennial, can thus exert a retarding influence on ground-water movement. It should not be regarded as an impermeable material, however, but rather as a confining material of low but finite permeability (an aquitard, rather than an aquiclude). Fracture systems, solution channels, and other discontinuities can provide significant additional hydraulic conductivity to a frozen soil or rock mass. The presence of segregated ice, as layers or lenses, may decrease the hydraulic conductivity to the point where the ground becomes effectively impermeable.

Depending on its composition, permafrost will therefore have a regionally and locally-variable, restrictive influence on both the spatial distribution and the rates of recharge, movement, and discharge of ground water.

4.1.3. Effects on Water Chemistry

4.1.3.1. Low Temperatures

Reaction and dissolution rates are reduced under the seasonal and perennial low-temperature conditions prevailing in the permafrost region. The solubilities of calcite and dolomite are somewhat increased because of the increased solubility of carbon dioxide at lower temperatures. The solubility of gypsum also increases slightly with decreasing temperature.

4.1.3.2. Permafrost

The presence of permafrost probably plays only a minor direct role in the development of any particular chemical composition of associated ground water. However, reduced rates of recharge, movement, and discharge tend to provide longer residence times for reactions between ground-water and aquifer materials than in non-permafrost areas.

Ranges of physical and chemical parameters for ground water in the permafrost region (Yukon and N.W.T.), shown in Table 4.1, are similar to those in non-permafrost areas.

	Yukon		N.W.T.	
Parameter	**Max.**	**Min.**	**Max.**	**Min.**
Flow, $L \cdot s^{-1}$	11,200	3.8	345	0.1
Temp., °C	54.5	0.0	64	-2.9
Cond., $\mu mhos \cdot cm^{-1}$	4,500	26	33,700	33
pH, units	9.5	3.1	9.3	2.9
Eh, millivolt	452	64	683	-252
Ca	630	1.3	1,823	1.2
Mg	374	< .05	1,190	0.5
Sr	14.6	< .05	14.9	< .007
Na	150	0.2	27,100	< .1
K	9.6	< .1	110	< .1
Li	0.18	0.01	1.7	< .005
Fe	450	0.04	2,600	< .001
Mn	6.7	0.01	62	< .001
Cu	0.19	< .001	0.28	< .001
Pb	0.1	0.004	0.19	< .001
Zn	29	< .001	1.0	< .001
As	-	-	0.014	< .0005
HCO_3	512	4.0	2,856	9.4
CO_3	22.8	12	151	8.5
SO_4	4,160	1.1	9,000	2.0
Cl	140	< .1	44,300	< .1
F	9	0.05	6.0	0.05
NO_3	7.6	0.01	55	< .003
PO_4	0.014	0.003	2.1	< .003
SiO_2	98	0.7	109	0.7
SUM	5,089	17.7	76,651	27.7

Table 4.1. Ranges of physical and chemical parameters for ground waters in the Yukon and Northwest Territories.

4.2. AQUIFERS IN PERMAFROST AREAS

4.2.1. Position Relative to Permafrost

In describing the hydrogeology of the permafrost region it is expedient to group aquifers under three major headings (Figure 4.3):

(a) suprapermafrost aquifers, occurring above the permafrost, with permafrost generally serving as a relatively impermeable basal boundary;

(b) intrapermafrost aquifers, occupying unfrozen zones or taliks (Canada, National Research Council of Canada, 1988) in the permafrost; and

(c) subpermafrost aquifers, occurring below the permafrost, with permafrost serving as a relatively impermeable upper boundary.

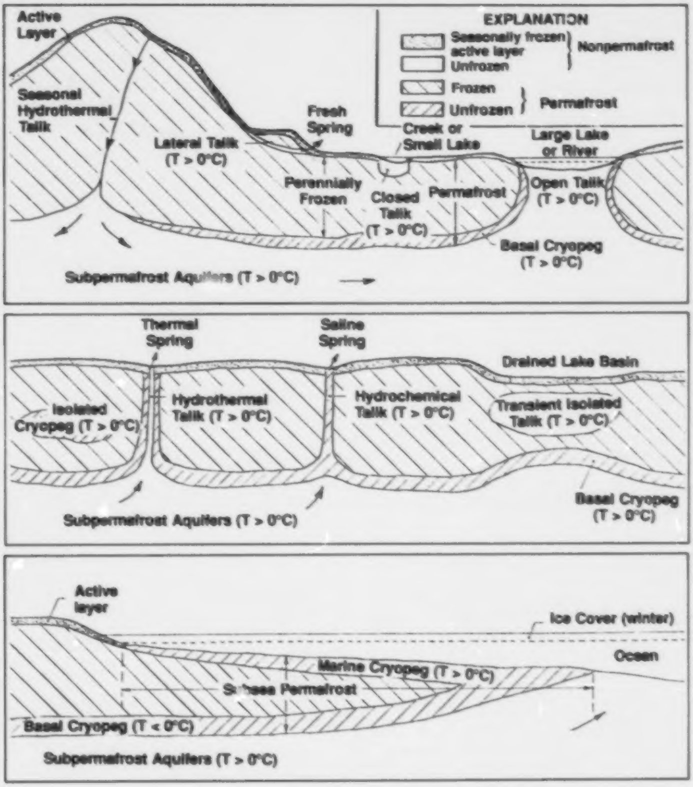

Active Layer
EXPLANATION
Seasonally frozen active layer
Unfrozen
Nonpermafrost
Frozen
Unfrozen
Permafrost
Seasonal Hydrothermal Talik
Fresh Spring
Creek or Small Lake
Large Lake or River
Lateral Talik (T > 0°C)
Perennially Frozen
Closed Talik (T > 0°C)
Permafrost
Open Talik (T > 0°C)
Basal Cryopeg (T > 0°C)
Subpermafrost Aquifers (T > 0°C)
Thermal Spring
Saline Spring
Drained Lake Basin
Isolated Cryopeg (T > 0°C)
Hydrothermal Talik (T > 0°C)
Hydrochemical Talik (T > 0°C)
Transient Isolated Talik (T > 0°C)
Basal Cryopeg (T > 0°C)
Subpermafrost Aquifers (T > 0°C)
Active layer
Ice Cover (winter)
Ocean
Marine Cryopeg (T > 0°C)
Subsea Permafrost
Basal Cryopeg (T < 0°C)
Subpermafrost Aquifers (T > 0°C)

Figure 4.3. Aquifers in permafrost areas: suprapermafrost aquifers in the active layer and closed river/lake taliks; intrapermafrost aquifers in open, lateral, and transient/isolated taliks, and isolated, basal, and marine cryopegs *(van Everdingen, 1987)*.

Various interconnections can exist among the three groups, as well as between them and the surface water in lakes and streams. Each of the three types of aquifers may be found in bedrock as well as in unconsolidated deposits.

4.2.2. Suprapermafrost Aquifers

Suprapermafrost aquifers can be divided into three subtypes on the basis of their freezing regime:

(i) aquifers in the active layer, undergoing seasonal freezing;

(ii) aquifers partly in and partly below the active layer, undergoing only partial seasonal freezing; and

(iii) aquifers below the active layer, but above permafrost, not subject to seasonal freezing.

4.2.2.1. Subtype I Aquifers

The depth of thaw and, therefore, the possible maximum thickness of subtype I aquifers, increases from north (< 1 m) to south (≥ 4 m). The thickness of active-layer aquifers may be further increased where the flow of water transports significant amounts of heat. By the time the active layer reaches its maximum thickness (near the end of the summer), seasonal freezing will start reducing it again from the surface downward.

4.2.2.2. Subtype II and III Aquifers

Such aquifers occur in depressions (closed taliks; see Figure 4.3) in the permafrost surface. High flow rates, resulting from a combination of high permeability and ample water supply, transport enough heat to prevent freezing of soil or rock over some depth interval below the active layer.

Subtype II aquifers are found in shallow closed taliks in alluvial and talus fans, in lake depressions, and in river terraces and valleys. Subtype III aquifers are associated with more extensive and deeper taliks existing beneath lake beds and river channels, and with taliks maintained by the discharge of thermal water. They are also found in the zone of discontinuous permafrost where the permafrost table may be well below the depth of seasonal freezing; and along marine coastlines where saline water can form "cryopegs" (hydrochemical taliks with temperatures below 0°C). The occurrence of subtype II and III aquifers is thus strongly dependent on the local distribution of permafrost, and on morphological and hydrological conditions.

4.2.2.3. Chemical Composition of Suprapermafrost Water

The chemical composition of suprapermafrost water generally reflects the influence of rainfall, snowmelt, and surface run-off. It is often high in organic (humic acid) content derived from muskeg areas. Where suprapermafrost water incorporates discharge from intra- and subpermafrost aquifers, its dissolved-solids concentration can be extremely high.

4.2.3. Intrapermafrost Aquifers

Intrapermafrost aquifers are not subject to seasonal freezing, and their extent is relatively constant, being affected primarily by long-term climatic (temperature) changes. During cooling trends the extent of intrapermafrost taliks can be expected to be reduced by aggrading permafrost, whereas during warming trends the reverse can be expected. The thawing induced by a long-term warming trend could reactivate "dormant" reserves of water; this would be reflected initially by increased run-off.

Intrapermafrost aquifers are found in open or "through" taliks, unfrozen zones that completely penetrate the permafrost; in "lateral" taliks, unfrozen layers in the permafrost; and in "isolated" taliks, unfrozen zones completely surrounded by permafrost (Figure 4.3).

4.2.3.1. Open Taliks

Unfrozen zones can occupy extensive areas beneath large lakes and major rivers. Their size and number decrease with decreasing size of the surface-water bodies, and with increasing latitude; eventually, such open taliks give way to closed taliks that gradually decrease further in size. Other open taliks are maintained through convective transport of heat by upward flow of warm (or hot) and/or mineralized subpermafrost water (hydrothermal taliks with temperatures above 0°C; hydrochemical taliks with temperatures below 0°C; van Everdingen, 1976a). Some open taliks are maintained by downward flow of surface or suprapermafrost water, e.g., in karst areas. Individual open taliks beneath rivers in mountainous areas may alternately carry upward-moving subpermafrost water, or downward-moving suprapermafrost (or surface) water, depending on the season.

4.2.3.2. Lateral Taliks

Unfrozen layers are sometimes found in permafrost where coarse- and fine-grained deposits alternate. The temperature in the more permeable layers is kept above freezing by the heat content of the water moving through the talik, supplied by either supra- or subpermafrost horizons. Such lateral taliks are of the hydrothermal type.

In the plains portion of large sedimentary basins, the basal interval of the permafrost is often unfrozen, due to strong mineralization. Such negative-temperature lateral taliks of the hydrochemical type are also know as "basal cryopegs" (Figure 4.3).

4.2.3.3. Isolated Taliks

Unfrozen zones entirely surrounded by permafrost may be of the hydrochemical type or they may be hydrothermal and transient (Figure 4.3). Isolated taliks of the hydrochemical type, also known as isolated cryopegs, have temperatures below 0°C; freezing is prevented by strong mineralization of the water. Such unfrozen salt-water zones in permafrost along seacoasts are called marine cryopegs (Figure 4.3).

Transient isolated taliks of the hydrothermal type develop beneath the beds of recently-drained lakes, where permafrost is actively aggrading, from the surface downward, into what were originally closed lake taliks (Figure 4.3). Their size decreases with time. They contain what was originally suprapermafrost water of subtype II and/or III, which undergoes a gradual increase in mineralization due to expulsion of salts from the ice phase during freezing. High pressures developing in such shrinking taliks can produce artesian discharge (Figure 4.4); they are the prime cause of the formation of pingos (Mackay, 1978, 1983).

4.2.3.4. Chemical Composition of Intrapermafrost Water

The chemical composition of intrapermafrost water may be similar to suprapermafrost water (in taliks with downward flow), or similar to subpermafrost water (in taliks with upward flow); cryopegs may contain saline or salt water, or brine.

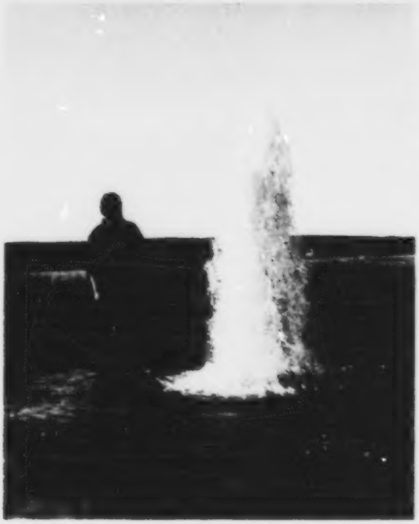


Figure 4.4.
Artesian discharge from the transient isolated talik below a pingo on the Tuktoyaktuk Peninsula, N.W.T. *(Mackay, 1978).*

4.2.4. Subpermafrost Aquifers

Subpermafrost aquifers, comprising all water-bearing materials below the permafrost, have temperatures equal to or higher than 0°C. In the following discussion a distinction is made between subpermafrost aquifers in unconsolidated deposits, in sedimentary terrain in general, in karst terrain, and in igneous and metamorphic terrain, because the effects of permafrost on the performance of subpermafrost aquifers are strongly dependent on the geological character of the aquifers.

4.2.4.1. Unconsolidated Deposits

Subpermafrost aquifers can be found in unconsolidated deposits wherever the thickness of such deposits exceeds the thickness of the permafrost. Aquifers in valley or terrace alluvium may receive recharge from precipitation or surface run-off directly, particularly if adjoining upland slopes do not contain permafrost. They may also receive recharge from above the permafrost via intrapermafrost aquifers.

4.2.4.2. Sedimentary Bedrock

Sedimentary strata of a variety of ages and lithologies form subpermafrost aquifers in sedimentary basins, in mountain ranges, and in intermontane depressions. Recharge to such aquifers is often severely restricted, particularly in the zone of continuous permafrost. The alternation of aquifers with aquitards and aquicludes can make many open taliks relatively ineffective as recharge points for deeper subpermafrost aquifers. Recharge along the margins of artesian basins may also be restricted by the presence of permafrost.

4.2.4.3. Karst

Karst development can create major subpermafrost aquifers in areas with soluble rock strata (carbonates, sulphates, or chlorides). Examples of such aquifers are found in the British Mountains in the Yukon, and in the Mackenzie and Franklin Mountains and the region west and northwest of Great Bear Lake in the Northwest Territories (van Everdingen, 1981). Karst aquifers are recharged through exposures of fractured rock, and through solution and collapse sinkholes. Small rivers may flow into such sinkholes (e.g., several streams in the Great Bear Lake region). Temporary lakes and ponds form in some of the sinkhole depressions every year during the snowmelt, when the rate of input exceeds the capacity of the subsurface drainage system (Figure 4.5); complete draining of such lakes may take until September, particularly if heavy rainfall occurs during the summer (Brook, 1983; van Everdingen, 1981). Karst recharge can account for 15 to 30 percent of the annual precipitation, particularly in areas where much of it falls as snow, recharging the karst system during snowmelt and thus escaping summer evaporation (van Everdingen, 1981).

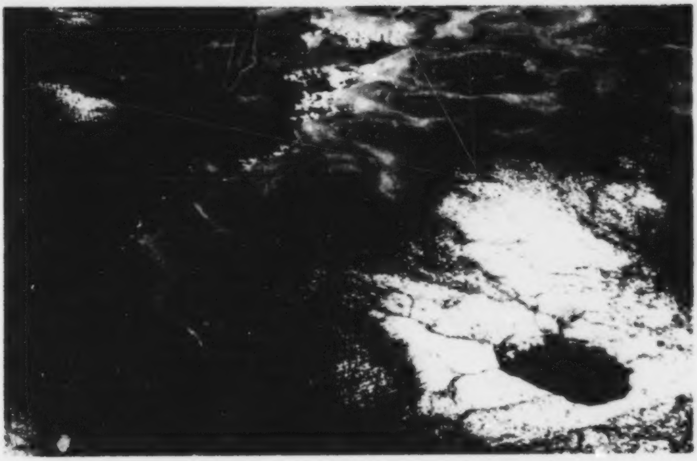
4.2.4.4. Igneous and Metamorphic Terrain

Igneous and metamorphic rocks can serve as subpermafrost aquifers, but their significance is limited by the vertical extent of the near-surface zone of fracturing relative to the permafrost. Where fracturing is widespread and extends well below the base of the permafrost, the rock mass may constitute a subpermafrost aquifer similar to aquifers in fractured sedimentary bedrock. Intersections of fracture zones with fault zones may, as usual, create more favourable conditions. Extrusive igneous rocks (lava flows in particular) may constitute excellent subpermafrost aquifers.

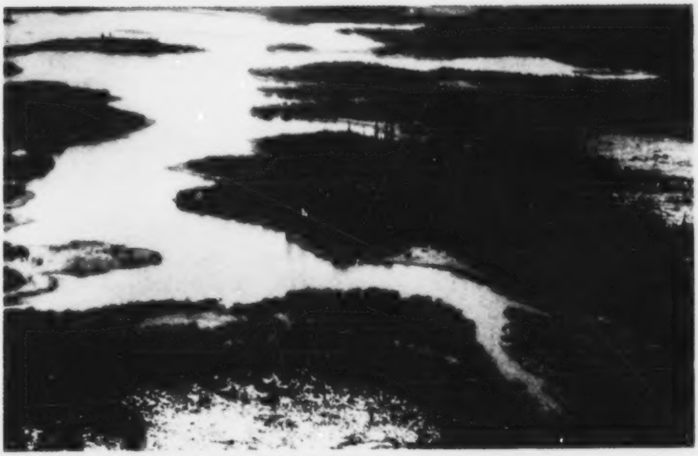
4.2.4.5. Chemical Composition of Subpermafrost Water

The chemical composition of subpermafrost water can range from freshwater of the Ca(Mg)-HCO_3 type, through brackish sulphurous and saline water, to sodium chloride or calcium/sodium chloride brines. The composition of subpermafrost water will depend primarily on the residence time of the water in the subsurface, and on the mineral composition of the aquifers. Low dissolved-solids contents are usually associated with fast flow systems in karst and in fractured, non-soluble bedrock.

In areas where permafrost is degrading, the water immediately beneath the permafrost, derived from melting of the original ice content, may be unusually low in dissolved-solids content. Low solution rates at low temperature, and low solubility of the carbonates in the absence of free CO_2, tend to limit re-dissolution of the carbonates and sulphates that were precipitated during the formation of the permafrost.



A Drained depression showing bedrock exposed around individual sink-holes (10 September 1975).



B Flooded depression (7 August 1976).

Figure 4.5. Large sinkhole depression in area of collapsed karst west of Smith Arm, Great Bear Lake, N.W.T. *(van Everdingen, 1981).*

Note contrast between vegetation zones along the edges of the depression, and vegetation on the surrounding permafrost.

4.3. GROUND-WATER HYDROLOGY

4.3.1. Comparison with non-Permafrost Areas

The occurrence of ground water, and the directions and rates of ground-water movement in the permafrost region are in general controlled by the same physical parameters as in areas without permafrost. The seasonal and areal distribution of precipitation determine the amount of water available for ground-water movement under the influence of gravity; and the distribution of rock types and structures in the subsurface determines the hydraulic conductivity and storage capacity, as well as locally preferred flow directions.

Mean annual precipitation in the permafrost region ranges from between 500 and 600 mm per year in the southern portion of the zone of discontinuous permafrost, to less than 100 mm per year over parts of the Arctic Islands (Figure 1.5). The amount of water available for recharge is therefore restricted by low precipitation over large portions of the region.

In addition, ground-water regimes in permafrost are seasonally affected by the formation of an aquitard in the active layer (for up to 10 months per year), and continuously by the presence of perennially-frozen ground. Thawing and subsequent recharge through the active layer may only occur during as little as two months each year. Although recharge can take place through frozen ground, the resulting recharge rates are usually one or more orders of magnitude lower than those for the same ground in unfrozen condition (Kane *et al.*, 1978; Kane and Stein, 1983a).

4.3.2. Infiltration and Ground-Water Recharge

Restriction of infiltration and recharge by permafrost is reflected by the presence of extensive muskegs (peatlands) and numerous lakes and ponds in the generally water-deficient region (Figure 1.10). However, many of the larger lakes and rivers have taliks of variable horizontal and vertical extent which present opportunities for either recharge to, or discharge from, deeper aquifers (Figure 4.3).

Recharge is least affected by permafrost in areas covered with coarse-grained unconsolidated deposits, and in areas where bedrock exposures present relatively large conduits (open fractures, interconnected joint systems, and solution channels for the infiltration of water. In crystalline rocks, where the continuity of fracture systems with depth is limited, recharge is more restricted. Recharge is very limited in areas covered with thick sequences of fine-grained unconsolidated deposits (e.g., lacustrine silts and clays), whose permeability may be even further reduced by the presence of ice lenses.

The concentration of recharge in smaller areas and at fewer points, by the presence of permafrost, can lead to very-high, local recharge rates. As an example, inflow into individual collapse-karst sinkholes in permafrost west of Great Bear Lake was found to exceed 1.0 $m^3 \cdot s^{-1}$ during the snowmelt period (van Everdingen, 1981).

4.3.3. Lateral Movement of Ground Water

In the permafrost region, lateral movement of ground water is confined to: (1) seasonal flow in the active layer during the frost-free season; (2) flow in taliks in the permafrost (e.g., river taliks); and (3) flow in subpermafrost aquifers (Figure 4.3). Information on this portion of northern ground-water flow systems is almost non-existent.

4.3.4. Discharge of Ground Water

The existence of active ground-water flow systems in the permafrost region is indicated by natural ground-water discharge phenomena similar to those observed in temperate regions. In addition, the presence of open-water reaches in ice-covered rivers (Figure 4.6), the formation of icings (Figure 4.7), and the presence of certain types of frost mounds (Figure 4.8) indicate the occurrence of ground-water discharge (Bryan, 1973; van Everdingen 1974, 1975, 1976b, 1982a, 1982b, 1988). Some of these discharge phenomena can be utilized to derive quantitative information on ground-water flow.

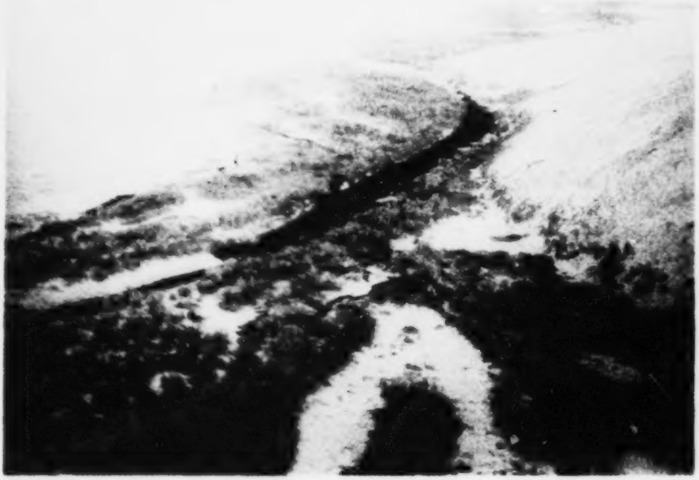


Figure 4.6. Open water maintained during winter by ground-water discharge in Babbage River, northern Yukon. Note contrast in vegetation between the discharge area (willows and poplars) and the surrounding tundra *(photo by R.A. Mutch).*

4.3.4.1. Springs

The discharge rates of springs and seeps can provide reliable information on available ground-water flow. The temperature and chemical composition of spring water often provide clues regarding the source of the water, i.e., supra-, intra-, or subpermafrost (Michel, 1977; van Everdingen *et al.*, 1979).

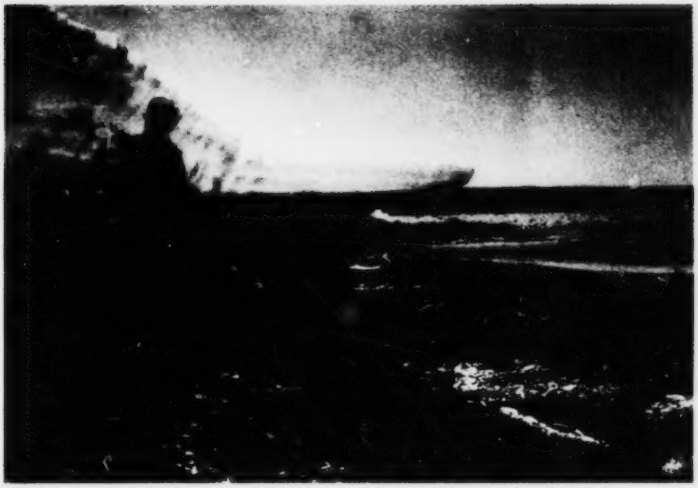


Figure 4.7. Icing on the flood plain of Crow River, northern Yukon, showing layering of ice and development of a thermo-erosional niche.

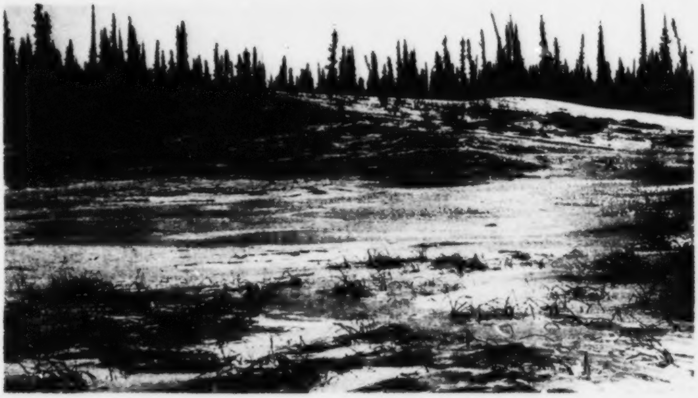


Figure 4.8. Frost mound at the Bear Rock discharge area near Fort Norman, N.W.T., 20 March 1976 *(from van Everdingen, 1982a).*

Perennial springs commonly discharge water from subpermafrost aquifers or, in some cases, from intrapermafrost aquifers. All thermal springs (T>10°C), most springs with discharge rates exceeding 5 $L \cdot s^{-1}$, and most mineral springs (with dissolved-solids concentration of 1 $g \cdot L^{-1}$ or more), derive their water from subpermafrost aquifers via open taliks. Seasonal springs and seeps that dry up shortly after freeze-up discharge water from suprapermafrost aquifers and, less commonly, from intrapermafrost aquifers.

The chemical composition of spring waters is usually a reliable indicator of the quality of the available ground water.

4.3.4.2. Base Flow

The base flow in streams without lake storage in their drainage basins represents primarily discharge of ground water. An example is provided by Figure 4.9, showing the streamflow record for Snag Creek (Yukon) at the Alaska Highway crossing (van Everdingen, 1988).

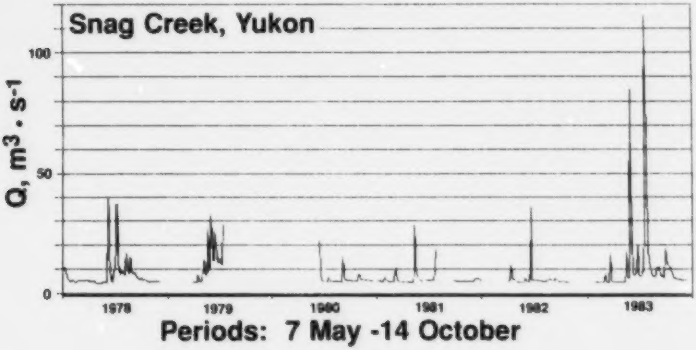


Figure 4.9. Daily discharge at the Snag Creek (Yukon) gauging station, for the period 1978-1983 *(from van Everdingen, 1988).*

As in temperate regions, a first approximation of the magnitude of the ground-water component in the surface run-off can be derived from the base-flow characteristics and from the spatial and seasonal variations in dissolved-solids concentration in the river water (Brandon, 1965; Hitchon *et al.*, 1969; Reeder *et al.*, 1972; Schreier, 1979).

The spatial variations in river-water chemistry reflect local inflows of mineralized water; this is illustrated by Figure 4.10, for a small tributary of the Mackenzie River, 38 km southeast of Norman Wells (van Everdingen, 1974). Seasonal variations in river-water chemistry, reflecting varying contributions by snowmelt and rainfall run-off, by discharge from the active layer, and by discharge from intra- and subpermafrost aquifers, are illustrated by Figure 4.11 (van Everdingen, 1987).

The ground-water contribution to base flow in the zone of discontinuous permafrost generally ranges between 2.0 and 5.0 $L \cdot s^{-1} \cdot km^{-2}$ (Williams and van Everdingen, 1973). Larger values reflect discharge from lakes in the basin. In the zone of continuous permafrost, winter base flow in many basins approaches zero, because most or all of the limited ground-water discharge is stored in icings. For example, Babbage River and Firth River (northern Yukon), which have no measurable streamflow at their respective gauging stations for 5 or 6 months each year, both receive as much as 1.4 $m^3 \cdot s^{-1}$ from ground-water discharge, which forms major icings in upper Babbage River, upper Firth River, and Joe Creek (van Everdingen, 1974).

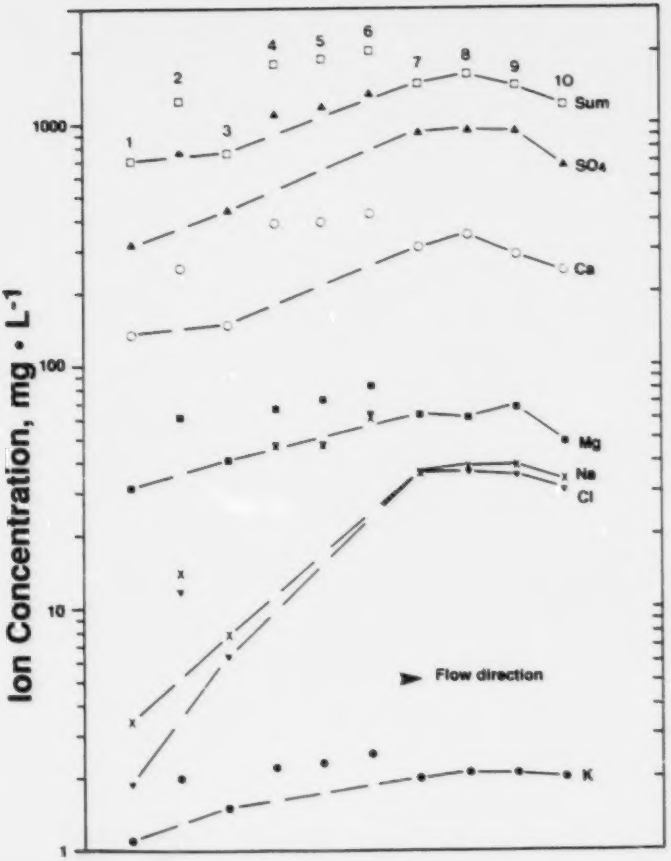


Figure 4.10. Chemical composition of spring waters (2,4,5,6) and variation in creek water-chemistry along the lower 11 km of Vermilion Creek, N.W.T. *(after van Everdingen, 1974)*.

During the winter, when base flow in many smaller rivers is derived exclusively from ground-water discharge, the quality of available ground water may be deduced from the chemical composition of the river waters. The chemical composition of ground-water discharge may of course change during the winter, due to changing proportions of water contributed by supra-, intra-, and subpermafrost sources. This may introduce some uncertainty into the interpretation of base-flow chemistry.

4.3.4.3. Icings

During the winter, much of the ground-water discharge in the permafrost region, which would normally maintain base flow in streams, may freeze

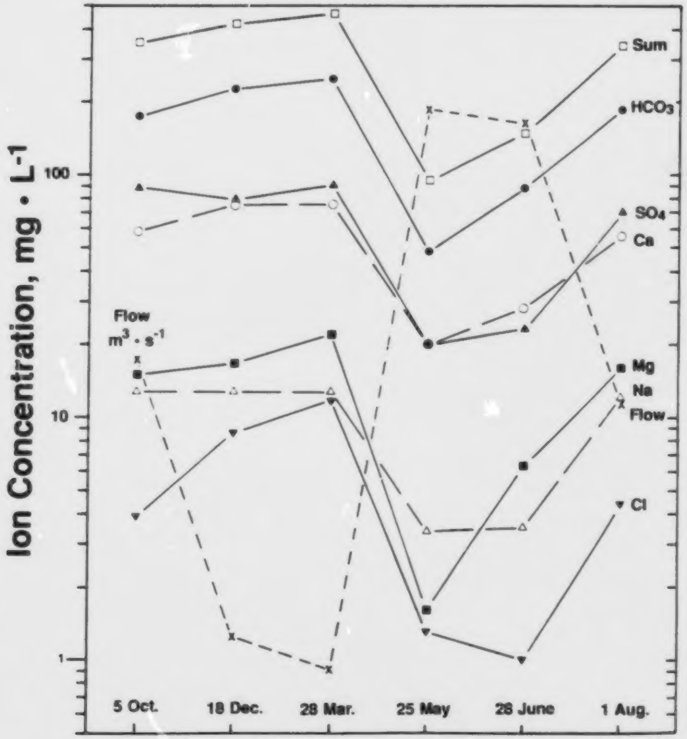


Figure 4.11. Seasonal changes in the chemical composition of the water of Ogilvie River, Yukon, below Engineer Creek *(data from Schreier, 1979).*

instead and form *icings* [general North American term], *aufeis* [widely used German term], or *naleds* [Russian term] (Sokolov, 1978). Icings can also form outside the permafrost region, particularly in mountain areas.

The distance between the point of discharge and the point where icing formation starts is a function of the relatively-constant discharge rate, temperature, and dissolved-solids content of the water; the geometry and gradient of the discharge channel(s); and the variable meteorological conditions of air temperature, humidity, and wind direction and speed.

Essentially all icings are to some extent related to the discharge of ground water. Icings fed by discharge of suprapermafrost water usually stop growing long before the end of the winter, when reserves are exhausted or the discharge conduits closed off by freezing. Icings fed by discharge of subpermafrost water, and some of those fed by intrapermafrost water, will keep growing till mean-daily air temperatures rise above 0°C (van Everdingen, 1982b).

Icings represent temporary above-ground storage of ground water discharged during the winter; the stored water is released by melting of the icing during the following spring and summer. Large icings may thus cause significant inter-seasonal redistribution of water resources. For instance, the icings developed over a period of about 6 months in the valleys of upper Babbage River, upper Firth River and Joe Creek take about 3 months to melt, increasing streamflow by 1.0 to 2.8 $m^3 \cdot s^{-1}$ during late spring and summer (van Everdingen, 1987). In general terms, the rate of run-off from melting icings appears to be 1.5 to about 4 times as high as the associated rate of ground-water discharge (Williams and van Everdingen, 1973); the actual ratio depends on the shape and exposure of individual icings, and on weather conditions.

4.3.4.4. Ground-Water Discharge Rates from Icing Volumes

Long-term studies in the U.S.S.R. have demonstrated that detailed information on the occurrence of icings can provide quantitative information on ground-water discharge rates, as an indication of ground-water resources of an area (Sokolov, 1978).

Measurements of the areal extent of individual icings (e.g., on air photos) can be converted into approximate volumes by the equation:

$$V_i = 0.96\ A_i^{1.09} \qquad [4.1]$$

where V_i is ice volume in 100 m^3, and A_i is icing area in 100 m^2. Icing volumes can also be determined in the field at the end of winter. If the duration of the freezing period or icing season, t_i, is known (e.g., from meteorological records), then the average ground-water discharge rate, Q_w, can be determined from:

$$Q_w = V_i/t_i \qquad [4.2]$$

This method should of course not be applied to icings that stop developing before the end of the winter, unless information is available on the actual duration of their growth period.

In the summer, after icings have melted, it is often still possible to determine their earlier presence and areal extent. The presence of an icing in a river valley during spring run-off may lead to accelerated lateral erosion by water channelling around the icing, and to braided channels. These characteristics assist in identifying icing areas during the summer. Mineral precipitates, formed during freezing of the discharged ground water, and left behind after melting of the ice, often persist for some time as a powdery coating on vegetation and rocks throughout the icing area (van Everdingen, 1974). The approximate extent of the icing can be determined from a careful survey of the distribution of the mineral precipitate.

4.3.4.5. Chemical Composition of Icings and Meltwater

The chemical composition of samples of icing ice is not a reliable indicator for the quality of the associated ground-water discharge, because of the uneven distribution of mineral precipitates and liquid inclusion in the icing (van Everdingen, 1974, 1982b). This uneven distribution has several causes. During the gradual freezing of the water flowing over and around an icing, dissolved-solids concentrations gradually increase in the remaining water, until saturation is reached with respect to the least-soluble minerals. Precipitation starts at that point and will continue, eventually involving more-soluble minerals. The last dissolved material is precipitated, or incorporated as brine inclusions, at the point where the last water freezes. The colder the weather, the closer to the discharge source this sequence of events take place. In addition, the continuously changing position of run-off channels on an icing will tend to distribute the water unevenly.

Meltwater from icings is also a poor indicator of the quality of associated ground water. During melting of the ice, the more soluble constituents (e.g., NaCl) go back into solution immediately, whereas calcite, dolomite, and gypsum re-dissolve much more slowly, forming a mineral slush on the surface and around the edges of a melting icing. The meltwater therefore generally has a lower mineral content than the associated ground water, with a higher ratio of chloride to bicarbonate and sulphate (van Everdingen, 1974, 1982b).

4.3.5. Importance of Ground Water

4.3.5.1. Fisheries

Open-water reaches, maintained during the winter in many northern rivers by ground-water discharge, play an important role in northern ecology, because they provide the only known overwintering and/or spawning areas for a number of fish species (Bryan, 1973).

4.3.5.2. Vegetation

Karst recharge helps maintain zones of non-permafrost vegetation around the edges of seasonally-flooded depressions (Figure 4.5). The zonation reflects on the one hand the relative duration of flooding (zones of sedges and grasses), and on the other hand the gradually decreasing thickness of the unfrozen layer (zones of willows, poplars, and white spruce).

North of the tree line, well-developed willows and poplars are found almost exclusively in ground-water discharge areas (Figure 4.6).

4.3.5.3. Water Supply

Data on the usage of ground water in the Yukon and Northwest Territories (covering a large portion of the permafrost region in Canada), presented by Hess (1986), are given in Table 4.2. More than 63 percent of the population of the Yukon relies on ground water for part or all of their water supply during part or all of the year (total ground-water use in 1981: 3.56 x 10^6 m^3). Only 0.6 percent of the population of the Northwest Territories relies on ground water (total ground-water use in 1981: 25,000 m^3); more than 9 million m^3 of ground water was extracted by the mining industry in the N.W.T. during mine dewatering. The difference between the two areas in actual usage of ground water reflects the much greater ground-water availability in the mountainous areas of the Yukon and the western portion of the District of Mackenzie, as compared to the low-relief areas in the District of Keewatin.

	Yukon	N.W.T.	TOTAL North of 60°N
Total Population	23,153	45,741	68,894
Ground-water Users[1]			
Municipal	13,700	202	13,902
Rural	975	80	1,037
Total	14,657	282	14,939
% of total population	63.3	0.6	21.7
Total Water Used[2]	15,928	19,012	34,940
Ground-water Use:			
Municipal	3,494	19	3,513
Rural	67	6	73
Manufacturing	0	0	0
Mining	0	9,168	9,168
Total	3,561	9,193	12,754
% of total water used	22.3	48.4	36.5

[1] Relying on ground water for part of all of their needs during part or all of the year.

[2] In $m^3 \cdot yr^{-1}$.

Table 4.2. Ground-water use in Canada north of 60°N *(data for 1981, from Hess, 1986).*

4.3.5.4. Recreation and Space Heating

Some 40 thermal springs with water temperatures ranging up to 64°C have been found in the mountainous areas of the Yukon and the District of Mackenzie (Crandall and Sadlier-Brown, 1976). Several of these springs discharge enough warm or hot water to support recreational developments (e.g., Takhini Hot Springs) and to satisfy local space-heating requirements.

4.4. HISTORICAL BACKGROUND

4.4.1. Pre-1970 Situation

Until recently, the ground-water hydrology of permafrost areas received relatively little attention in Canada. Before 1970, the only information available on the occurrence of ground water in the permafrost region in Canada consisted of observations on a few springs, and data on a small number of widely-scattered and commonly shallow wells (Beschel, 1963; Brandon, 1965). As late as 1978, the Hydrological Atlas of Canada indicated the existence of only one ground-water observation well in the permafrost region, located in northern Alberta (Canada, Fisheries and Environment, 1978).

4.4.2. Recent Advances

A variety of research projects were undertaken during the last 15 years in connection with planned and proposed developments, such as the Mackenzie and Dempster highways; the Mackenzie Valley, Alaska Highway, and Arctic Islands gas pipelines; the Norman Wells oil pipeline; and the Churchill/Nelson and James Bay hydro-power developments. In particular, the studies related to proposed highway and pipeline projects have contributed to the knowledge of ground-water occurrence in the region, and they have provided valuable additions to the available data base.

Ground-water discharge areas in the southern and central portions of the Mackenzie transportation corridor were described briefly by Hughes *et al.* (1973) and Rutter *et al.* (1973). The isotopic composition of spring waters in the area were discussed by Michel (1977) and van Everdingen *et al.* (1979). The isotopic composition of soil water and ground ice were studied along both the Mackenzie Valley and Dempster Highway corridors (Michel and Fritz, 1978; Michel, 1983). Thermal springs in northern B.C., the Yukon and the District of Mackenzie were described by Crandall and Sadlier-Brown (1976). Additional ground-water information has come from water-well surveys (Intera, 1975) and water-well drilling (e.g., Trimble *et al.*, 1983).

An initial set of 18 ground-water data maps (1:250,000) with notes, covering parts of the National Topographic Series (NTS) sheets 95, 96, 106, 116, and 117, was released into open file by the Task Force on Northern Oil Development (van Everdingen, 1974/75). A set of 34 route maps (1:50,000), showing springs, seepages, open-water areas and icings along the Alaska Highway pipeline route in the Yukon, were prepared for IWD, Pacific and Yukon Region (van Everdingen, 1978). A major ground-water discharge area along the proposed route north of Beaver Creek was described by van Everdingen (1988).

The distribution, formation, and melting of icings was studied with the use of satellite imagery (van Everdingen, 1975, 1976b) and time-lapse photography (van Everdingen, 1982b). Satellite imagery and time-lapse photography were also used in the study of seasonal flooding and draining of karst depressions in permafrost (van Everdingen, 1976b, 1981).

Analyses of the isotopic composition of dissolved sulphates in a number of sulphurous spring waters in the area were used to determine which of the evaporite-bearing stratigraphic units (the Upper Cambrian Saline River Formation or the Lower Devonian Bear Rock Formation) is the source of the sulphates (van Everdingen and Krouse, 1977; van Everdingen *et al.*, 1979, 1982). Similar analyses of acidic spring waters revealed that their acidity and sulphate contents are derived from oxidation of metal sulphides in the subsurface (van Everdingen *et al.*, 1985).

The role of ground-water discharge in the formation of frost mounds was investigated at a site near Fort Norman (N.W.T.) using time-lapse photography (van Everdingen. 1978, 1982a; van Everdingen and Banner, 1979); and at a site in North Fork Pass (Yukon), using specially designed piezometers (Pollard and French, 1983).

The occurrence of various types of unfrozen zones in permafrost was discussed, in connection with proposed new permafrost terminology, by van

Everdingen (1976a, 1985). A number of new terms are defined in the new "Glossary of Permafrost and Related Ground-Ice Terms" (Canada, National Research Council of Canada, 1988).

The potential interactions between pipelines and ground water in permafrost areas were reviewed by van Everdingen (1979).

Development and adaptation of special instrumentation and methods during this period included (1) camera systems for long-term automatic time-lapse photography (Banner and van Everdingen, 1979a); (2) frost gauges and freezing gauges to monitor seasonal freezing and thawing (Banner and van Everdingen, 1979b), and formation and degradation of permafrost at a pipeline test site (van Everdingen and Carlson, 1988); (3) time-domain reflectometry (TDR) for measurement of unfrozen-water content (Kane *et al.*, 1978; Kane and Stein, 1983a; Patterson and Smith, 1983; Smith and Tice, 1988); (4) electro-magnetic (EM) methods for mapping of permafrost (Hoekstra and McNeill, 1973; Sartorelli and French, 1982); and (5) ground-probing radar (GPR) to locate unfrozen zones in the subsurface (Lafleche *et al.*, 1988).

4.4.3. Agencies Involved in Recent Research

Groups involved in the above ground-water-related studies in northern Canada included:

1. National Hydrology Research Institute (Environment Canada) and its predecessors;
2. Fisheries and Marine Service (Environment Canada);
3. Water Quality Branch (Environment Canada);
4. Geological Survey of Canada (EMR);
5. Northern Roads Directorate (Public Works Canada);
6. Water Resources Division (DIAND);
7. Aquatic Environments Ltd. and Northern Engineering Services Ltd., both under contract to the proponents of the Mackenzie Valley gas pipeline;
8. Foothills Pipelines (Yukon) Ltd.;
9. Department of Earth Sciences, University of Waterloo;
10. Department of Geography, University of Ottawa;
11. Department of Geography, University of British Columbia;
12. Geotechnical Science Laboratories, Carleton University;
13. Engineering and environmental consulting companies.

4.4.4. Incentives for Recent Research

The major incentives for the above ground-water-related studies in northern Canada were:

1. Geotechnical concerns (stability of proposed highways; effects of culvert icings; stability of pipeline crossings of rivers; frost heave).
2. Environmental concerns (protection of permafrost terrain against accelerated erosion; protection of fish overwintering and spawning area; provision of adequate culverts for fish passage; protection of wildlife).
3. Water-resources concerns (protection of water resources against suspended and dissolved contaminants).
4. Professional interest of the scientists involved played a significant role in the initiation of a number of the above research projects.

4.5. CURRENT RESEARCH

Although ground-water-related northern research efforts are considerably reduced, a number of studies continue, and others have been added.

1. Studies of the potential effects of climatic change on permafrost. Examples can be found in Smith and Riseborough (1983) and Etkin *et al*. (1988), and in the program for Special Sessions A and B held on 6 February 1989 during the 5th International Cold Regions Engineering Specialty Conference in St. Paul, Minnesota, organized by the American Society of Civil Engineers. Although these studies do not deal directly with ground-water hydrology, their findings will provide a basis for the prediction of potential effects of climatic change on northern ground-water hydrology (van Everdingen, 1987; see also Section 1.6.2.3.).
2. Investigation of seepage from mine-tailings impoundments, using thermometry and ground-probing radar (e.g., Lupin gold mine). Agencies involved include the Geological Survey of Canada, mining companies, and engineering/environmental consultants.
3. Provision of alternative ground-water supplies for settlements that depend on unreliable surface sources; and provision of thermal ground water for addition to cold-water supplies, to prevent freezing of distribution lines. Agencies involved include the Yukon Department of Public Works, the Geological Survey of Canada, and engineering/environmental consultants.
4. Application of TDR for the determination and monitoring of unfrozen-water contents in frozen ground (e.g., Carleton University).
5. Development of methods for measurement of hydraulic conductivity in a permafrost environment, and laboratory studies of mass and heat transport (revived initiative by the National Hydrology Research Institute).

4.6. FUTURE RESEARCH NEEDS

4.6.1. Major Unknowns

Major unknowns with respect to existing research approaches in the study of northern ground-water hydrology include:

1. Hydrogeological properties of intra- and subpermafrost aquifers, including permeabilities, storage capacities, piezometric levels, flow velocities, and residence times.
2. Magnitude of ground-water recharge through permafrost in fractured and karstic rocks.
3. Basin-wide magnitude of ground-water recharge.
4. Reliable and environmentally-safe methods for long-term recording of water-table and piezometric levels.

4.6.2. Subjects Requiring Research

In the following listing of subjects requiring research, no order of priority is implied. Classification of the research efforts by type is indicated as follows: **A** = study of air photos and/or satellite imagery; **D** = studies using available data; **F** = field studies; **L** = laboratory studies; **M** = physical or mathematical modelling.

1. Regional study of the chemistry of both surface and ground waters, and of the seasonally varying influence of ground-water chemistry on the composition of surface waters, to obtain regional estimates of available ground-water resources (**D/F/M**). The initial data phase would require the verification of physical, chemical and location data for the more than 500 ground-water analyses produced during recent research projects north of 60°N.
2. Regional study of the distribution, regime and magnitude of icings, to evaluate available ground-water resources (**D/A/F/M**).
3. Effects of major icings on channel hydraulics, flow rates, and water chemistry of their host streams (**F/L/M**).
4. Problems of aquifer and well testing in permafrost: (a) extent of heating required to prevent freezing of water in wells and piezometers; (b) extent of insulation required to prevent thawing of permafrost around pumped and flowing wells and heated observation wells and piezometers; and (c) effects of frozen boundaries and progressive thawing of partially-frozen layers on water-level drawdown and recovery (**F/L/M**).
5. Problems of producing water wells in permafrost areas: (a) the extent of heating and/or insulation required to prevent freezing of standing water during well shut-downs; (b) the amount of insulation required to prevent thawing of permafrost around producing wells (due to conductive heat-loss from the water) which could affect the integrity of the well casings (**F/L/M**).

6. Long-term monitoring of ground-water levels in selected basins with discontinuous and continuous permafrost (**F**).
7. Effects of climatic change. How would gradual warming or cooling of the climate affect the rates of ground-water recharge and discharge; the availability and quality of ground-water resources; the overwintering and spawning areas for northern fish species; and the icing activity and icing problems along northern highways (**F/M**)?
8. Modification of ground-water (and soil-moisture) movement to minimize icing problems and frost-heave damage (**M/F**).
9. Extent to which freezing will prevent subsurface migration of contaminants from mine tailings (**F/L/M**).
10. Effects on ground-water hydrology from thawing of permafrost in and around mines (**M/F**).
11. Effects of mine dewatering on thermal and ground-water regimes (**M/F**).
12. Effects of high embankments (highways, railways, dams) on the hydraulic conductivity of water-bearing closed taliks beneath small streams, and on the incidence of icings (**F/M**).
13. Extent to which freezing will prevent subsurface migration of contaminants from saline drilling fluids at abandoned drilling sites (**F/L/M**).
14. Effects of various types of backfill in pipeline trenches on ground-water movement in the active layer (**M/F**).
15. Several of the above studies would benefit from further development of surface geophysical methods for the delineation of the frozen ground that forms seasonal and perennial aquitards and/or aquicludes (**L/F**).
16. Some of those studies will also require the development or adaptation of low-cost instrumentation for the measurement (or long-term automatic recording) of total and unfrozen-water contents in soil or rock, and of changes in phase, water level or pressure, and water chemistry, in situations where the sensing elements will be subjected to extreme pressures from seasonal or incidental freezing (**L/F**).
17. The study of northern ground-water hydrology, and of the potential interactions between permafrost, engineering developments, and hydrological processes would also benefit from the establishment of a central data base or clearing-house for the exchange, compilation, and reporting of all types of northern hydrological data (**D**).

4.6.3. Future Incentives for Ground-Water Research

Future incentives for studies of northern ground-water hydrology will likely be the following:

1. Accidents or emergencies involving existing domestic, municipal, or industrial ground-water supplies in northern Canada.

2. Requirements for additional or alternative water supplies for expanding settlements and industrial developments.
3. Prevention or abatement of ground-water pollution in areas where ground water currently provides water supplies, or where ground-water pollution can be expected to lead to contamination of surface waters (e.g., around tailings impoundments).
4. Confirmation of the suspected climatic warming trend, which can be expected to affect northern ground-water hydrology.
5. Re-activation of proposed northern highway and pipeline projects that can be expected to affect and be affected by ground-water hydrology.

4.6.4. Major Impediments

The major impediments for the sophisticated research needed in the field of northern ground-water hydrology will be:

1. The high cost of northern field operations and specialized field instrumentation, to be financed from limited budgets. Future research efforts will therefore likely have to be concentrated in areas near settlements or major developments, which may not provide the best choice of conditions from a scientific point of view.
2. Lack of research specialists with experience in northern ground-water investigations. ❄

5

Hydrology of Floating Ice

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5.1. INTRODUCTION

An obvious and notable feature of lakes and rivers in the North is that they are ice-covered for portions of the year. Indeed if the occasional formation of an ice cover is taken as a definition of the North (Section 1.2.1.3.) then it would encompass most of the land mass of the Northern Hemisphere including the northern United States (Figure 1.1). A significant portion of floating-ice hydrological research has been conducted south of 60°N, but many of the results obtained from these more temperate latitudes are highly relevant to conditions north of 60°N. Hence, this chapter includes discussion of research undertaken south of 60°N in Canada and the U.S.A., mainly to provide a more complete picture of the hydrology of floating ice.

It is the ice cover on inland waters that is referred to here as floating ice. Its hydrological influence arises through its effect on the flow and water level in a stream, the water level in a lake, and through the seasonal storage represented by the ice itself, the snowcover it carries, and the channel and lake storage it induces. This hydrological significance of floating ice can be substantial. Indeed it can be argued the hydrological extremes of common interest - floods and low flows - are as much a function of stream processes, through the action of ice, as they are of the catchment processes of traditional concern. While the peak discharge is primarily a function of catchment processes such as snowmelt, the peak water level - the cause of flooding - is very much a function of the ice conditions on the stream. This is particularly so for the North where the snowmelt peak is the peak discharge of the year and can occur while the stream is still ice-covered or otherwise influenced by ice in the channel. For example, in the period 1983-87 ice jams were involved in some 30% of the flood events across Canada (Canada, Environment Canada, 1988a). In New Brunswick ice-jam floods such as that shown in Figure 5.1, are responsible for more flood damage than open-water floods (Kindervater, 1985): the 1987 ice jams on the St. John River alone caused $30 million damages (Humes and Dublin, 1988). On the other side of the country, in northwestern Canada, the flood threat at almost all riverside communities is primarily due to ice jams, not summer floods. Figure 5.2, which shows a comparison between ice-jam and open-water flood stages for the Hay River at Hay River, Northwest Territories, illustrates the situation.

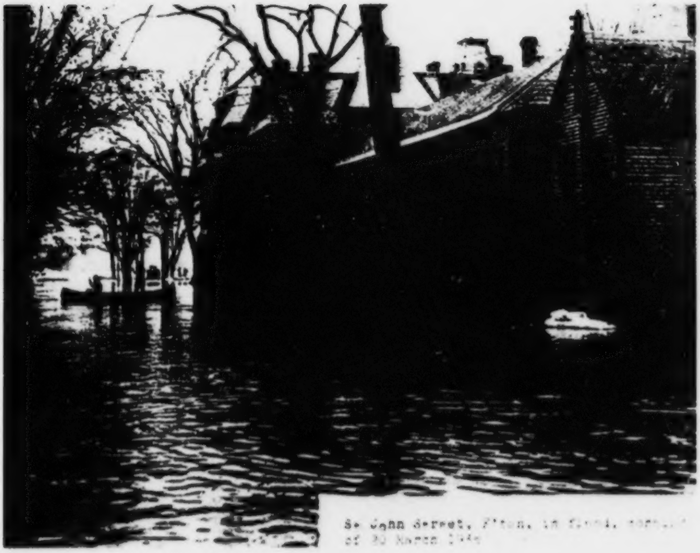


Figure 5.1. Ice-jam flood, Fredericton, New Brunswick, 1936 *(from Kindervater, 1985).*

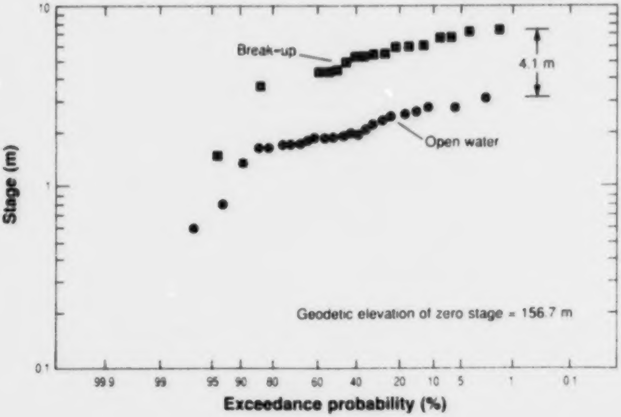


Figure 5.2. Exceedence probability distributions for ice jams and summer floods, Hay River at Hay River, N.W.T. *(after Gerard and Stanley, 1988b).*

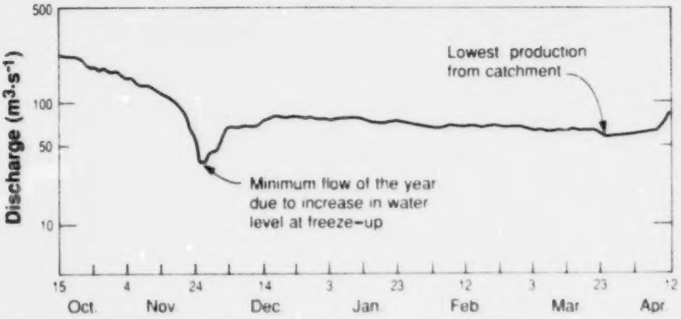


Figure 5.3. Winter hydrograph of the Clearwater River at Draper, Alberta, 1976-77 *(Water Survey of Canada data).*

At the other extreme, low flow at a site on a cold-region stream can also depend heavily on ice processes. A striking illustration of this is provided by the fact that the discharge over Niagara Falls was halted on March 29, 1848 by ice obstructing the outlet of Lake Erie! A more common circumstance is shown in Figure 5.3: the minimum discharge shown occurred in October, due to ice formation upstream, rather than in late winter when discharge from the catchment was a minimum. Figure 5.4 shows the low flow frequency curves for several rivers in northern Alberta. The marked 'abnormality' in the curves for the smaller streams is explained by ice effects (Gerard, 1981).

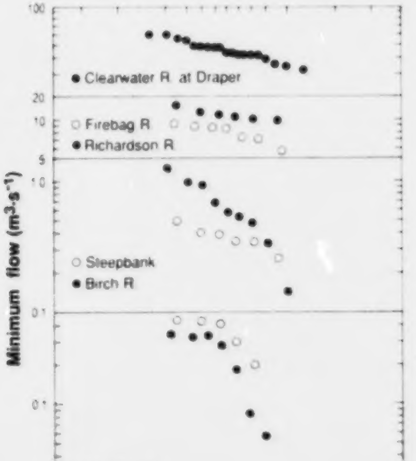


Figure 5.4. Minimum flow probability distributions for streams of various sizes in northern Alberta *(from Gerard, 1981).*

As well as influencing the extremes, ice effects can have a major influence on the winter hydrograph of cold-region streams in general. In streams the volume of water stored as ice, and as channel storage due to the increase in water level caused by the ice, can represent a significant portion of winter flow which does not become available until spring. Although little investigated, this may be particularly so for the lake-dominated rivers of the Canadian Shield where slight changes in the resistance to flow from the outlet due to changes in the ice cover can trigger enormous changes in lake storage. An extreme example of this is provided by the Great Lakes (see, for example, Derecki and Quinn, 1986). On the other hand snowfall on lake ice can cause an increase in flow from a lake (Adams, 1981). The weight of water displaced from the lake must equal the weight of the snowfall on the lake ice (if the latter is simply floating, with little restraint from the shore, as is usually the case). Hence a 0.3 m snowfall will displace approximate 30 mm of water from the lake, a flow that can be very significant in a stream in mid-winter in a catchment with a large proportion of lakes. Therefore, unlike on land, a water equivalent of snow falling on lake ice is made immediately available as flow (while a similar amount will be made available in the spring when the snow melts, it should not be counted twice when evaluating the catchment yield). Similarly, in fall, snow falling on a still-open lake or river runs off immediately whereas that falling on land can remain until spring (Williams, 1971). For small northern lakes the duration of the ice cover and its thickness (as it affects the volume of water remaining under the ice cover) are very important considerations for aquatic fauna, with winter kill always being a possibility. Water supply reservoirs and waste-water treatment ponds must also be designed for the volume of storage lost to ice formation each winter. Adams (1981) discusses the influence of ice types on the duration of a lake [and river] ice cover as well as a host of other hydrology-related aspects of ice-covered lakes.

The hydrological effects of floating ice are therefore of major concern in cold regions. As indicated, they are a major cause of floods in Canada, but these floods are not just significant because of the damages and loss of life they may cause. In other circumstances they can be beneficial. For example, the multitude of lakes in the vast and environmentally important Mackenzie and Peace-Athabasca Deltas in western Canada depend on the periodic flooding caused by ice jams (Figure 5.5) to refill and refresh them (Marsh, 1986).

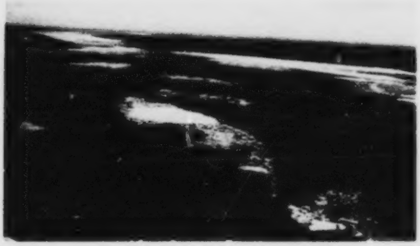


Figure 5.5.
Mackenzie Delta during break-up showing the ice-jam flooding of the delta lakes that is so important to the wildlife of the Delta.

And there are other concerns. Water intakes have to be designed to be below the low stages generated by ice in the fall (Bruce and Clark, 1966; Gerard, 1981) while the pumps and ancillary equipment must be above the flood stages of ice jams in the spring. Hydro-electric power operations can be severely restricted by the need to avoid releasing discharges so high as to trigger ice-jam floods downstream or releasing discharges so low that, in the presence of ice effects, the discharge will fall below the regulatory minimum. For example, B.C. Hydro sustains a 'significant loss of revenue' when operations have to be curtailed to avoid freeze-up flooding along the Peace River, while Manitoba Hydro estimates it costs more than $1,000,000 per year to release sufficient flow from South Indian Lake to avoid total freeze-off of the Churchill River (Foulds, 1988).

Given this obvious significance of 'floating ice hydrology', it is astounding to note the lack of interest shown in the subject by the hydrological community. A recent review of snow and ice hydrology (Collins, 1987b) mentions the topic of ice on inland waters almost as an aside, while the remainder of the review is devoted to glacial hydrology. A similar situation pertains to the chapter on ice in the tome on hydrology by Chow (1964). More disconcerting is the almost complete absence of discussion of the subject in hydrological texts. Even a major text that purports a Canadian orientation (Gray, 1970) does not deal with ice jam floods or other aspects of ice hydrology. It is therefore not surprising that Canadian graduates are largely ignorant of this important aspect of their hydrology.

It was not always so. In the early years of Canada's development ice effects were a major preoccupation of the scientific and engineering community. For example, the first paper in the Transactions of the Canadian Society of Civil Engineers was on frazil ice (Henshaw, 1887); a major presentation to the Royal Society of Canada by its President was on the subject of ice and floods (Keefer, 1898); and the first major text on ice engineering was published by a Canadian physicist (Barnes, 1928). There were likely many reasons for this interest, one of which is shown in Figure 5.6: one of the major ice hydrology problems occurred regularly on the doorstep of Canada's largest city.

After these early years Canadian interest in research in floating ice hydrology appears to have waned, with some notable exceptions. Certainly there has been nothing to match the attention devoted to the subject in the U.S.S.R. However, over the past two decades the subject has undergone a revival and

Figure 5.6.
Ice jam flood,
St. Lawrence River at
Montréal
(from Keefer, 1898).

begun to regain its rightful place in the spectrum of research in the natural sciences. A major step forward was the publication on the winter regime of lakes and rivers by Michel (1971). Since then much of the credit for the revival internationally is due to the series of Ice Symposia organized by the International Association for Hydraulic Research. Within Canada it is largely due to the activities of the National Research Council Associate Committees. One important result of the consequent advance in the state of the art is that ice jams are finally beginning to become a routine consideration in flood studies in Canada (Ontario Ministry of Natural Resources, 1985; Watt, 1989). But there is much yet to be done.

In the following presentation the status of floating ice hydrology is briefly reviewed. In doing this the emphasis has been placed on the hydrological aspects of floating ice, as against the hydraulics, mechanics and thermal aspects, although they are all closely connected. The presentation also concentrates on hydrology in its narrow engineering sense, being taken to refer to the influence of floating ice on flow and water levels in inland waters. In particular, discussion is confined to the four features of most hydrological significance. First, the circumstances that generate high stages, which are important in flood studies, the operation of hydro-electric plants and the design of riverine structures. Second, low discharges, the hydrological feature of most environmental significance and which is again important in hydro-electric plant design and operation, and in designing water supplies and water intakes. While floating ice has little effect on long-term yield it can be responsible for major shifts in the timing of the yield within a year through its influence on storage, so this is the third consideration. Finally, a fundamental resource for hydrological planning, design and operation is accurate information on discharges. This is based on flow measurements. The ice cover can have a significant and often detrimental effect on these measurements so these are also considered briefly. Geographically the emphasis is on activities in Canada, with reference to work in other countries only when similar work is lacking in Canada, or when it provides a useful contrast to the Canadian effort. Within these confines the coverage was further restricted to avoid overlap with other chapters in this book. In particular, the hydrological influence of aufeis and glacial lake outburst floods (jøkulhlaups) are described in Section 4.3.4.3. and 6.5.3.2. respectively. It should be noted, however, that the interaction between these ice-related circumstances can be important. For examples, the effects of a glacial lake outburst flood in mid-winter have been exacerbated by the presence of an ice cover on the release stream (Bowers, 1978); and aufeis formation can be the result of a floating ice cover freezing to the bed.

To define the bounds of the effects of floating ice on inland waters, and to provide a background for the later discussion, the processes of ice formation and demise on inland waters are briefly reviewed.

5.2. ICE PROCESSES ON RIVERS AND LAKES: AN OVERVIEW

Freeze-up of a small well-mixed lake in calm weather occurs in a straightforward manner. When the lake has cooled sufficiently that the surface water temperature falls to a little below zero during the diurnal minimum, a thin and fragile ice sheet will form over the lake surface. It will remain for the rest of the winter if it thickens fast enough to resist disturbance by wind, or melt by a subsequent thaw.

On a large lake the process is much less well defined. The first water to cool sufficiently for ice to form is that in the shallow protected areas around the shore. Here sheet ice - or border ice - forms in much the same manner as on a small lake. However in the centre of the lake the surface is continually disturbed by wind. This prevents formation of the initial thin sheet and instead the heat demand from the cooling atmosphere is satisfied by the formation and growth of small (typically a millimetre or so in diameter) fine discs of ice called frazil. Being ice, the particles are less dense than water and, although they can be entrained and carried to considerable depth by the wind-generated turbulence, they eventually accumulate at the surface to form frazil slush. Here they are exposed to the cold air temperatures. This, along with the wind agitation, causes the frazil slush to gradually agglomerate to form frazil pans. With air temperatures cold enough, the concentration of pans, slush, and sheets broken from the shore ice, can eventually become high enough to dampen the surface disturbance by the wind and finally allow an initial solid ice cover to form over the lake surface. This ice cover will then thicken by ice growth on the underside of the ice cover. The ice so formed is coarse-grained, with a distinct columnar structure, and is almost transparent. The growth of this clear ice will be terminated by a snowfall of sufficient weight to 'sink' the ice cover. Typically an accumulation about half the ice thickness is sufficient to do this. Then water can flood out over the ice, saturating the lower layers of snow. This saturated snow begins to freeze, forming snow ice. This ice is fine-grained and, due to the fine air bubbles incorporated into it, is white and opaque. If another snowfall occurs before the saturated snow layer is completely frozen, an unfrozen slush layer will be left behind as the process is repeated. The presence of the saturated snow renders the ice cover isothermal after a short time, so there will be no further heat transfer from the bottom of the ice cover and ice growth there will cease until all the saturated snow layers are frozen.

As explained by Williams (1971) and Adams (1981), because of the different grain structure and vastly different transparency to solar radiation, the relative proportions of clear and white ice in a snowcover has considerable significance for the life of the cover and events in the water beneath it. While a large proportion of clear ice can be expected on northern lakes where the initial ice thickens rapidly and there is little snowfall, the ice cover on lakes in regions of heavy snowfall can be predominantly white. However, as pointed out by Adams and Prowse (1981), even on the same lake the proportions of clear and white ice can vary significantly, primarily due to the effects of snow drifting. As the white ice is formed from snow, its presence must be considered when conducting a snow survey for a catchment with a significant proportion of lakes and rivers.

While the water temperature at the under-ice surface in a lake is at freezing, that just below can be significantly above freezing due to the winter 'inversion' caused by the fact that water reaches its maximum density at 4°C. Because of this 'warm' water within the lake, the flow at the outlet of the lake is above freezing. The outlet can therefore remain open long after the remainder of the lake is ice covered, as shown in Figure 5.7. This can have significant repercussions on the variation in flow from the lake, and the winter hydrology of the outlet stream.

The situation at freeze-up in a river is somewhat similar to that of a large lake, with two major differences: the turbulence in a river is generated by its own flow, and is therefore ever-present except in pools above rapids, bars, weirs or dams. It is sufficient to prevent any thermal stratification of the flow so that the water temperature remains within a few hundredths of a degree throughout the flow depth. Again the first ice to form is sheet ice over the quiet water of the shallows along the banks. Out in the central region of the stream, the flow and turbulence is usually sufficient to prevent the formation of sheet ice on the surface. Instead frazil forms and is mixed throughout the depth.

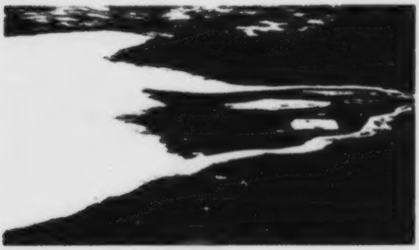


Figure 5.7.
Outlet of Lake Laberge, Yukon *(from Carmack* et al., *1987).*

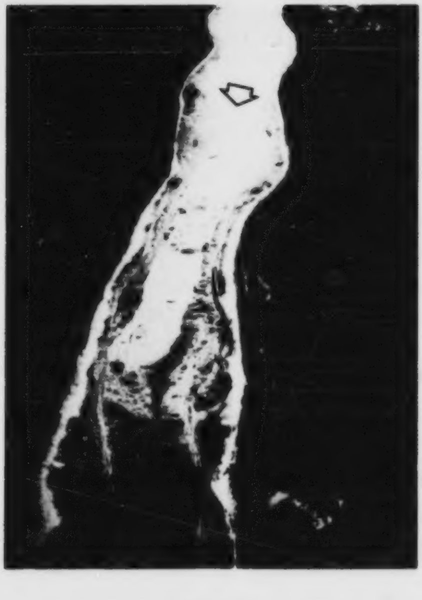


Figure 5.8.
Lodgement, North Saskatchewan River near Myrnam, Alberta.

When first formed this frazil is actively growing and has a propensity to attach itself to anything it touches. This includes the bed material where frazil will accumulate if it is large enough to resist the buoyant force of the accumulation. This is known as anchor ice. Such deposits can blanket rapids and gravel bed rivers to a thickness of the order of a metre. In a shallow stream, these deposits can grow to project through the water surface to freeze solid and directly obstruct the flow. The deposits are then known as ice dams.

If the turbulence generated by the flow is not too intense, frazil particles will accumulate at the surface to form slush, which in turn will agglomerate into frazil pans. Hence, as the cold weather continues, border ice grows out from shore, while the concentration of frazil pans increases on the remaining free surface. Eventually, at some point along the river, the remaining space between the border ice edges will be insufficient for the passage of the frazil pans. Then the pans will lodge between the border ice, and other pans moving down from upstream will begin to accumulate behind the lodgement, as illustrated in Figure 5.8. This initial pack gradually extends upstream as more frazil pans arrive, forming the initial ice cover on the river. Downstream of the lodgement the production of frazil and the growth of frazil pans must begin again, which delays freeze-over downstream.

As the ice cover develops on the water surface the resistance to flow in the remaining waterway under the ice increases. The water level must therefore rise to provide a waterway sufficient to carry the discharge with the increased resistance. This depth increase is typically greater than 30% of the mean depth of the open-water situation. Furthermore, the ice cover or pack floats with more than 90% of its thickness submerged. This necessitates a further increase in water level to provide the required waterway. The nature of the initial pack that forms depends on weather conditions during its formation and on the discharge and geometry of the river. If the river is steep and fast, and the air temperatures mild during formation, the initial pack will be thick and the rate of frazil-pan production slow. In these circumstances the increase in water level as the ice cover passes a section will be large but the pack progression slow. For example, on the Nelson River, a stream characterized by alternating pools and rapids, the initial pack must develop thicknesses of up to 12 m to allow the pack to progress over the many stretches of fast water (Newbury, 1968). Because of the enormous volume of ice required for progression of a pack of such thickness, the progression is slow, despite bitterly cold air temperatures during freeze-up. On the other hand, if the river is flat, the flow slow and the cold intense during formation of the pack, the pack will be thin and frazil production high. In these circumstances the pack that forms will be simply one pan in thickness and the progression upstream rapid.

The variation in water level as the pack passes can have considerable hydrological significance. As shown in Figure 5.9, the water required to fill the channel storage associated with the increase in water level is abstracted from the flow moving into the reach from upstream. This means

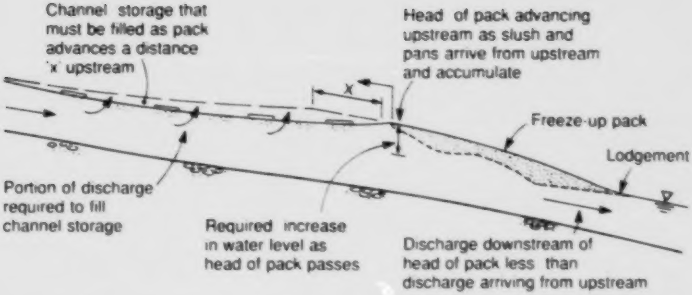

Channel storage that must be filled as pack advances a distance 'x' upstream
Head of pack advancing upstream as slush and pans arrive from upstream and accumulate
x
Freeze-up pack
Lodgement
Portion of discharge required to fill channel storage
Required increase in water level as head of pack passes
Discharge downstream of head of pack less than discharge arriving from upstream

Figure 5.9. Loss of discharge to channel storage during progression of an ice cover.

the discharge moving downstream below the pack is less. This effect can be responsible for the minimum discharge of the year, as shown in Figure 5.3.

After the pack is in place, it will begin to freeze solid. The ice so formed is fine-grained and a little translucent. If the pack is thin enough, and the season long and cold enough, the freezing front will progress through the pack into the flow underneath. The ice that is formed then is coarse-grained and transparent, as described for lake ice. Also as for lake ice, the freezing of the pack, or growth of clear ice under it, will be terminated by a snow accumulation heavy enough to submerge the ice surface.

If open water exists upstream, frazil will be generated and carried under the ice cover to be deposited to form a slush-ice layer. As the freezing front moves through this, fine-grained ice is formed that is somewhat translucent and often has a smoky hue due to sediment particles incorporated in the ice.

The above processes continue throughout the winter. Because the ice cover is initiated at different times in different places, and its character depends on the local stream characteristics, it can be expected the final ice cover will be quite heterogeneous, possibly varying from a relatively thick ice cover that consists totally of frozen frazil slush due to a thick initial pack at one location, to thin, totally clear ice over a polynya (an area of open water left in the original pack) at another location. If there are reaches of fast water, areas of warm ground water or lake outflow, or a large polynya, there can also be areas of open water that survive the winter.

With the onset of warm weather, the temperature of the ice cover increases until melt begins at the bottom and, if the ice is exposed to solar radiation, within the ice cover itself. This initial melt within the cover occurs at the grain boundaries, a process known as candling, and which, particularly in clear ice, with its columnar structure, can cause a dramatic reduction in the ice-cover strength while having little effect on its thickness. Being fine-grained and opaque, white ice resists the debilitating effects of solar radiation on its strength with the result that areas of the ice cover that are predominantly white ice can remain strong when areas of clear ice, such as in a freeze-up polynya, have decayed completely.

At about the same time, snowmelt begins and causes an increase in water level. The bottom-fast ice along the shore is not free to follow the increase in water level and becomes flooded, exposing the water to warming by solar radiation and air temperature. Any open-water areas expand as the heat the water absorbs melts the ice cover downstream. Areas of thin ice become weak and either melt or, if at the upstream end of an open-water area, break away due to the drag of the current, to accumulate at the downstream end of the open water.

If the discharge remains reasonably steady these processes of melt, decay and fragmentation will continue until the river reach is free of ice. However, if the discharge increases substantially, or an accumulation of fragmented ice develops to the extent that the load it exerts on the downstream ice cover causes the ice cover to fail, an ice run can be initiated. As the ice moves the depth falls, releasing water from channel storage, the reverse of what happened at freeze-up. This releases a surge, or river wave that causes an increase in discharge and water level downstream and helps sustain the ice run. Eventually this ice run will stall, either because it has been attenuated to the point that it cannot break the in-place ice cover, or because it encounters a sudden increase in waterway width or depth, such as at the confluence with a reservoir, lake, or larger main stream, or at a sudden reduction in channel slope, that absorbs the surge maintaining the ice run. When the ice run stalls an ice jam has formed and the water level will increase substantially. Eventually the ice jam will fail or move, possibly releasing another surge that will trigger an ice run again if any ice remains downstream. This process is repeated, not necessarily sequentially, until the whole river is finally free of ice.

On a lake the process of ice decay and melt begins as on a river. It is typical that the ice closer to the shore melts first (Williams, 1966; Woo and Heron, 1989), releasing the main body of ice to be moved about the lake by the wind. As on a river, the rate of melt, and particularly decay, of a lake ice cover depends on the proportion of clear and white ice. Indeed, as described by Adams and Prowse (1981), this effect can be sufficient to reverse the shore-to-centre progression of melt mentioned above.

On a large lake, wind can assist break-up by blowing large ice floes about the lake once they have been freed from shore by melt. However, on more moderate-sized lakes the ice more-or-less decays and melts in place, only being disturbed by wind when it is in a very frail state.

The above events are typical of a truly cold region, so that the water body experiences only one freeze-up and one break-up each year. In more temperate regions there may be more than one freeze-up and break-up cycle in a given year, whereas in other years there may be none at all. In such situations events become a strong function of the quantity of ice that can be generated in each cold spell. In North America such a situation is typical of the Maritimes, southern Ontario and New England, and of British Columbia and the northern Pacific States of the U.S.A. Inland and north of these locations the former scenario is more typical. On lakes in the High Arctic the situation can be such that there may be no break-up at all in a particular year.

It was made clear earlier that floating ice plays a major hydrological role in northern areas. In what follows, the research that has been carried out on this aspect of hydrology is briefly reviewed and some suggestions are made for possible future directions the research might take. The emphasis is on the major research thrusts, by which is meant those that are reasonably long-term, systematic and non-proprietory. In common with the other chapters, the subject is first broached by providing some historical background on the research effort in Canada. Then the current research is described, with an indication of the reasons for the work. Finally some suggestions are made about future research needs. There is an unavoidable overlap between these various sections. The references have been selected to illustrate the state-of-research, not to present a complete literature review.

5.3. HISTORICAL BACKGROUND

5.3.1. Floods and High Stages

Because of its immediate engineering significance, the major research on the hydrological effects of floating ice has concerned ice jams (Figure 5.10) and their generation of high water levels. A recent review of the various aspects of this subject is given by Gerard (1988). The first real progress in understanding the processes and mechanics of ice jams seem to have come from work on log jams by Kennedy (1958) and from the laboratory and field work undertaken by Hydro-Québec (Cartier, 1960), Université Laval (Michel, 1965), and La Salle Hydraulics Laboratory in Quebec (Pariset and Hausser, 1961; Pariset, Hausser and Gagnon, 1966) and Manitoba Hydro (Newbury, 1968). These investigations were initiated because of the influence of ice on hydro-electric power plant operation, flooding and navigation. Since then other laboratory investigations at the University of Iowa (Uzuner and Kennedy, 1976) and major programs of field observations of break-up ice jams by the Alberta Research Council (Gerard, 1975; Beltaos, 1983; and Andres and Doyle, 1984), the Cold Regions Research and Engineering Laboratory (CRREL) of the U.S. Army Corps of Engineers (Calkins, 1983), the National Hydrology Research Institute (NHRI) (Prowse, 1986), the National Water Research Institute (NWRI) (Beltaos, 1988) and the University of Alberta (Rivard, Kemp and Gerard, 1984; Gerard and Stanley, 1988a) have largely confirmed the validity of these early analyses of ice jams.

The emphasis in these studies was on the so-called equilibrium section of a fully-developed ice jam - the section characterized by uniform waterway depth and pack thickness, and importantly, the largest depth along the jam (see Figure 5.10). It has been common practice to presume the water levels along an ice jam in a non-uniform reach could be simulated by assuming the pack is at equilibrium at each section along the pack. The agreement achieved by such an assumption has apparently been quite good, as illustrated by the analysis of ice jams that caused major flood damage on the Delaware River in the U.S.A. (Zufelt and Doe, 1986).

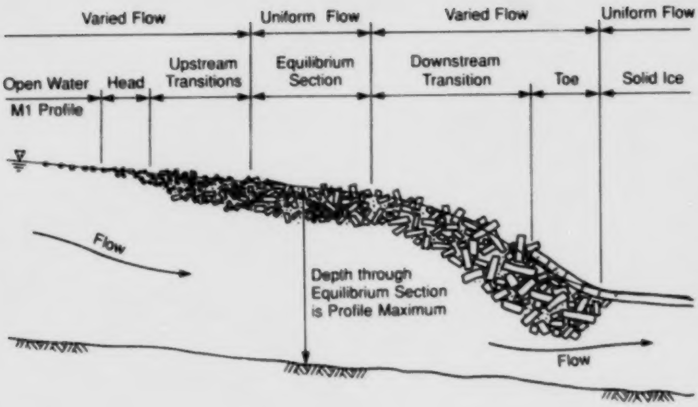
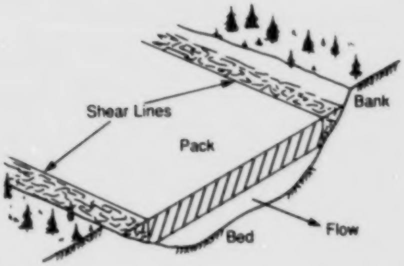

Varied Flow
Uniform Flow
Varied Flow
Uniform Flow
Open Water
M1 Profile
Head
Upstream Transitions
Equilibrium Section
Downstream Transition
Toe
Solid Ice
Flow
Depth through Equilibrium Section is Profile Maximum
Flow

Figure 5.10.
The presumed configuration of a fully-developed ice jam and photograph of a break-up ice jam on the Hay River at Hay River, N.W.T.


Shear Lines
Bank
Pack
Flow
Bed

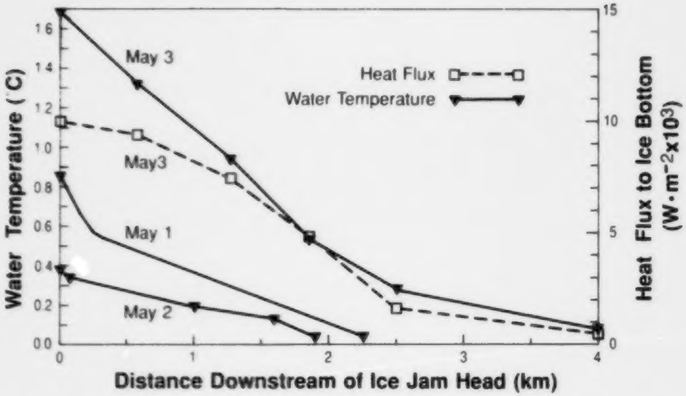


Figure 5.11. Variation in water temperature beneath an ice jam, Liard River at Fort Simpson, N.W.T. *(from Prowse and Marsh, 1989).*

With the understanding of the equilibrium section of ice jams on a reasonably firm footing, attention has shifted in recent years to other important features of an ice jam, especially the toe region. Prediction of the water level and pack thickness along the full length of an ice jam is not possible without first developing an understanding of this portion of the jam. It also presumably plays a vital role in defining the load applied by the jam to the downstream ice cover, and the bed and banks, and thereby in the formation and failure of the jam. And, it also has a major influence on the nature of the surge released by the ice jam when it fails. Flato and Gerard (1986) describe an algorithm that allows calculation of pack thickness and water level over the full length of the jam, including the toe.

Parkinson (1982), Prowse and Marsh (1989), and Prowse (1990) have demonstrated the major role melting can play in the life of an ice jam, and the latter investigators have made a significant contribution to defining the nature of the heat transfer from the inflowing water to the ice jam pack. An example of the water temperature decrease as it flows under the pack is shown in Figure 5.11; given the associated discharge, such a profile represents the melting of an enormous amount of ice.

When an ice jam fails it releases a surge that moves downstream causing the sudden increases in water levels and velocities that are an almost universal feature of the description of ice-jam floods. A beginning was made by Nuttall (1973) and Mercer and Cooper (1977) on analysing the nature and magnitude of such surges through analytical and numerical studies of the unsteady flow associated with potential ice-jam releases in the lower reaches of the Mackenzie River. An analytical parametric study, which included consideration of waves generated by ice-jam formation, was carried out by Henderson and Gerard (1981), based on a simple unsteady flow analysis. This was followed by another numerical analysis of a particular situation by Beltaos and Krishnappan (1982) and, together with some preliminary laboratory experiments, by Joliffe and Gerard (1982). Wong *et al.* (1985) describe similar laboratory experiments.

Major field studies of freeze-up jams and accumulations include the multitude of studies carried out for the La Grande River (Michel and Drouin, 1981; SEBJ, 1984), Archipel (St. Lawrence River) (see, for example, LeVan and Michel, 1986), Nelson River (Hopper *et al.* 1978) and Peace River (Keenhan, Panu and Kartha, 1980) hydro-electric projects. Other studies on the St. Lawrence River have been carried out by Clarkson University in New York (Shen and Ackermann, 1981), again the concern being the operation of the nearby hydro-electric plants. CRREL has also carried out studies of freeze-up jams as they are a common cause of flooding in the small steep streams of New England and the U.S. northwest (Cunningham and Calkins, 1984; Calkins, 1985). A recent appearance in the technical literature is the extensive work by the Chinese on freeze-up of the Yellow River (Sun and Shen, 1988).

The above studies have been concerned with the detailed mechanics of ice jams. When fully developed this understanding of ice jams will form the basis for an important component of an algorithm to predict the ice regime of a river. However, it can already be used to define an ice-jam rating curve for a site, which can then be utilized in the calculation of ice-jam flood levels for planning and flood damage reduction (Gerard and Calkins, 1984; Anderson *et al.*, 1988). However a problem with accounting for ice jams in such studies is the difficulty of assessing the frequency of ice-jam formation.

To avoid the 'frequency' problem, an alternate, and more common, approach is to analyse the historical ice-jam flood levels directly, if data is available, as is routinely done for open-water floods. Because ice-jam floods are very local and site specific, it is difficult to transpose information from another site. Generally, only information in the immediate vicinity of the site is useful and in many cases this is limited. In particular, it is unlikely there will be a hydrometric gauge in the reach of interest. It is therefore usual to have to extract the maximum amount of information on water levels and probabilities from the historical record for the particular site. A simple technique to do this, and typical case studies, have been described by Gerard and Karpuk (1979) and Stanley and Gerard (1988).

Another hydrological need related to ice jams is a procedure to predict ice-jam floods. The difficulty with this is the present inability to forecast the occurrence of an ice jam in any given year. In the past this problem has been approached empirically by relating the timing and magnitude of the peak stage during break-up to antecedent hydrometeorological circumstances. The most extensive applications of this approach have been in the U.S.S.R. and many examples are described by Shulyakovskii (1963). In North America, accounts of reasonably successful applications of this approach are those of Beltaos (1984) and Gerard and Stanley (1986).

5.3.2. Low Flow

Compared to ice-jam flood studies, concern with the effect of floating ice on low flows is relatively recent (Gerard, 1981). However, it can be argued that in many parts of the country it is perhaps a more important question than that of ice jams because of the almost universal and increasing significance of pollution and its effect on inland waters. As indicated earlier, low flows can be a major consideration for dam operations that are required to release sufficient water to be environmentally satisfactory, to provide dilution of pollutants and to ensure water for satisfactory operation of water intakes. Recent examples of major environmental concerns with low flows are the impact studies for the Susitna Project in Alaska (Paschke and Schoch, 1985; Paschke and Coleman, 1986), concern over the winter diversion from the Nechako River in B.C. (Blachut, 1988) and the string of pulp mills proposed for the Athabasca River in Alberta.

As described earlier, the progression of the freeze-up pack can cause a significant reduction in discharge downstream, a discharge that may well be the minimum discharge of the year. Rivers that receive a large portion of their winter flow from lakes are likely to be especially prone to the influence of ice because of the effect of an ice cover on the outflow from lakes, and the large change in the volume of water that must accompany even a small change in water level at the lake outlet. The interaction of lakes, ice and minimum flow is a largely neglected, but likely important, subject.

On small streams, such as is shown in Figure 5.12, the water stored in the ice cover itself can be a significant loss to the winter flow. Again an important variation on this occurs if the catchment has a large proportion of lakes (Adams, 1981).

Another possibility is that the stream or lake will freeze to the bed, isolating that portion of the drainage network from ground-water inflow and effectively reducing the catchment area for winter discharges and perhaps triggering the formation of aufeis. However, as pointed out by Gerard (1981), the presence of a lake of reasonable size on an otherwise small steam may serve to keep the stream from freezing to the bed and hence open to ground-water inflow. The movement of the frost line and its interaction with the base flow may explain the temperature-related variations in winter flow noted even when the air temperature remains well below freezing

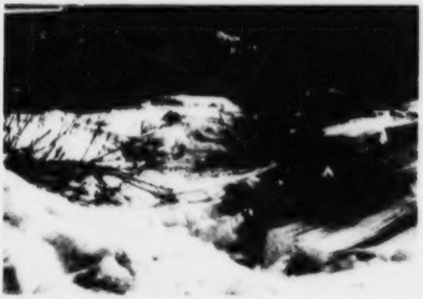


Figure 5.12.
Typical small stream in winter. A small flow continues under the ice cover.

(Chin, 1966). Surprisingly, these features have been little studied. However, Wankiewicz (1984) has made a start with a detailed investigation of the winter regime of a small stream reach in northern Canada.

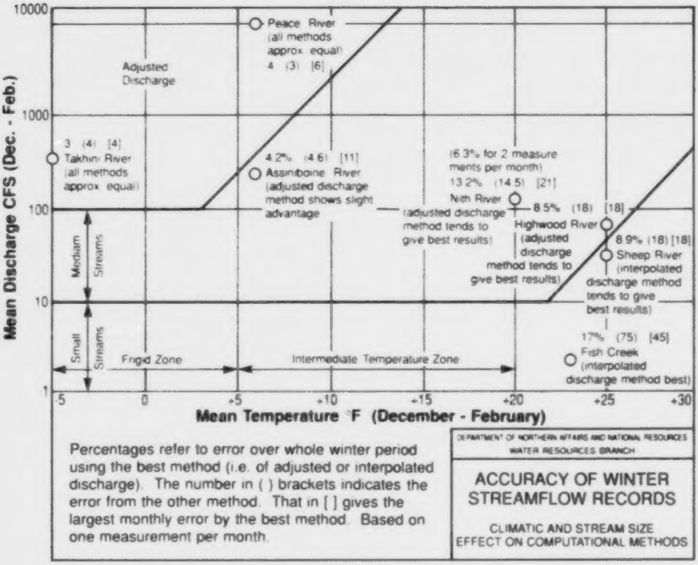
5.3.3. Flow Duration

Low flows and floods are the hydrological extremes of most common interest. However, there is also a need for continuous simulation of the discharge throughout the winter, primarily for hydro-electric operations. The needed computational facilities and a reasonably adequate understanding of ice processes, particularly outside of the break-up season, have existed for some time, yet it is only recently that serious attention has been directed to the development of non-proprietary versions of computer models of the winter regime of lakes and rivers, although proprietary versions have apparently been available for many years (see, for example, Petryk *et al.*, 1980). This is unfortunate because, even if not perfect, such models are a valuable research tool in that they provide documentation and demonstration of the state-of-the-art. For example, Calkins (1984) describes the content of such a model and illustrates how its application provides immediate indication of the weak links in the understanding of ice processes and of the required emphasis for future research. Huokuna (1988) describes a model developed for assessing the ice regime downstream of hydro-electric power plants in Finland. Destuynder (1986) outlines a model developed for assessing the influence of ice covers on reservoir water temperatures, and indicates that development has begun on a model of river-ice regime to assess its influence on the flow at the water intakes of nuclear power plants in France. Recently Shen *et al.* (1988) have published the background to a very comprehensive ice-regime model developed at Clarkson University.

A particular problem with continuous simulation is the need to define the break-up process. While thermal break-up is reasonably well understood and can be simulated satisfactorily (Andres, 1984) the same cannot be said of dynamic break-up. An early attempt to simulate dynamic break-up in a physical model was described by Michel and Abdelnour (1975). More recent efforts have been directed at developing an understanding of the process through numerical modelling and field observation (Ferrick *et al.*, 1986; Williamson, 1989). In particular the work by Ferrick and others has resulted in a recent breakthrough in the ability to forecast dynamic break-up and progression (Ferrick and Mulherin, 1989).

5.3.4. Flow Measurement

A fundamental tool in hydrology is flow measurement. As consideration of winter flows becomes more significant, so too does the measurement of these flows. In rivers of reasonable size accurate discharge measurement poses little problem (Pelletier, 1988), simply requiring a slight adaptation of summer techniques and the usual care with instrument calibration. However the same is not true of small streams. Here current metering under the ice in the shallow obscured flow is difficult, and installations to measure the flow, which may operate satisfactorily in the summer, become victims



Percentages refer to error over whole winter period using the best method (i.e. of adjusted or interpolated discharge). The number in ( ) brackets indicates the error from the other method. That in [ ] gives the largest monthly error by the best method. Based on one measurement per month.

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WATER RESOURCES BRANCH

ACCURACY OF WINTER
STREAMFLOW RECORDS

CLIMATIC AND STREAM SIZE
EFFECT ON COMPUTATIONAL METHODS

Figure 5.13. Dependence of errors in interpolated discharge during the winter season on climate, stream size and method *(from Rosenberg and Pentland, 1966).*

of ice formation. Reviews and guidelines for mitigating these problems have been provided by Jaffray (1975) and Rantz (1982) but many problems remain.

While the situation with regard to direct discharge measurement in other than small streams is reasonably satisfactory, the same cannot be said of the procedures used to interpolate the discharges in the stream over the period between the discharge measurements. This was the subject of an extensive investigation by the Water Survey of Canada (Rosenberg and Pentland, 1966). The main results of the investigation are summarized in Figure 5.13. The study indicated that the reliability of the various techniques depends on river size and winter severity.

For large rivers which have a gradual and steady decline in discharge over the winter and that have a stable ice cover form early in the season, any reasonable technique of interpolation seems to provide reliable results. Indeed it has been argued (Santeford and Alger, 1984b) that in such situations rational use could be made of a winter rating curve, if explicit account was taken of the variation in the nature of the ice cover at the gauge (not the discharge-measurement) site. Such an approach is simply automated and would relieve the many hours usually devoted to manual interpolation of the discharges. While this proposal is likely applicable in many cases and should certainly be pursued, particularly for those streams with a long stable ice-cover period, the somewhat irregular winter rating curve for the Takhini

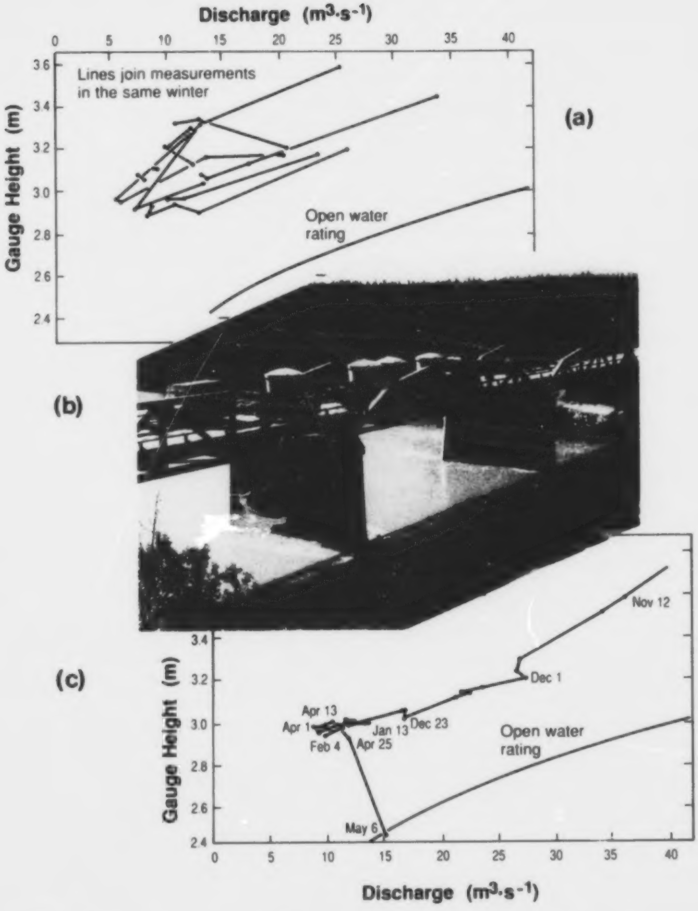

Figure 5.14: Open-water rating curve and winter discharge measurements over several years. Takhini River at the Alaska Highway, Yukon: (a) Measurements for various winters;
(b) Photograph of the Takhini River at the Klondike Highway crossing some 40 km downstream, June, 1989;
(c) Variation over the winter of 1965-66 *(from Chin, 1966).*

River shown in Figure 5.14 suggests care will likely still be required until variations in the freeze-up and break-up ice regime are better understood.

For a small stream, particularly in a more temperate climate where the ice cover will be intermittent through the winter period and for which the discharge is sensitive to ice effects, the situation is markedly different. Then any method is likely to be unreliable and, somewhat surprisingly, water-level measurements may not help the situation; indeed Figure 5.13 indicates they can add more noise than information.

A special situation arises in subarctic-nival regimes (Section 1.4.1.2.) where the peak flow of the year is due to snowmelt and coincides with break-up (Figure 5.15). In this circumstance the relation between stage and discharge is variable and difficult to define during the most crucial time of the water year. None of the standard techniques will work in such a situation. Furthermore, the peak run-off period is brief and the sites inaccessible, so that it is difficult to define even the open-water rating curve using the usual periodic visits. To overcome these problems, the Water Survey of Canada have instituted spring field camps when frequent discharge measurements are made to define the run-off hydrograph (Canada, Environment Canada, 1984b), in some cases using the moving-boat technique to allow discharge measurements to be made almost during the ice runs. Other novel techniques that might be used to estimate discharge in such circumstances are described by Prowse *et al.* (1987).

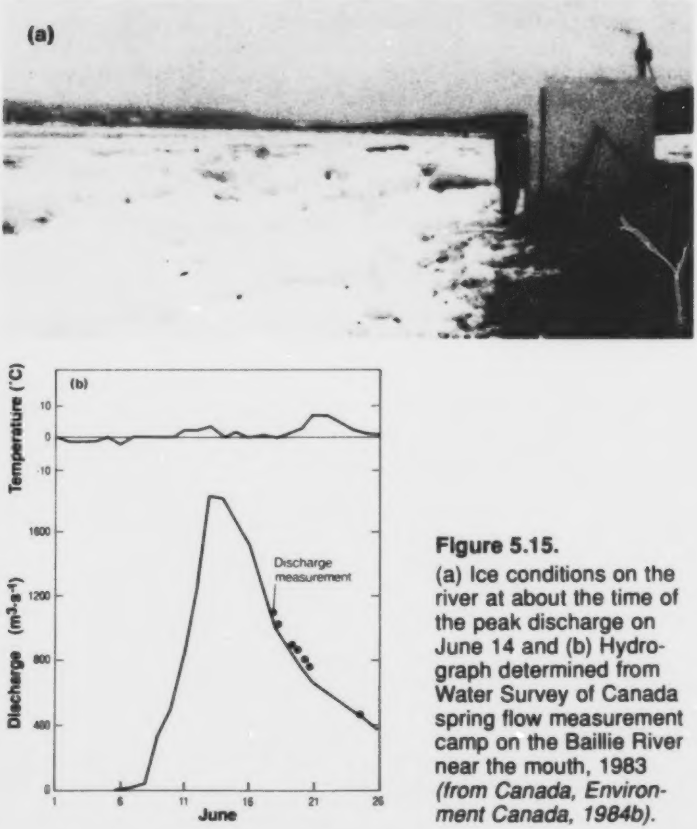

Figure 5.15.
(a) Ice conditions on the river at about the time of the peak discharge on June 14 and (b) Hydrograph determined from Water Survey of Canada spring flow measurement camp on the Baillie River near the mouth, 1983 *(from Canada, Environment Canada, 1984b).*

5.4. CURRENT RESEARCH

Research into floating-ice hydrology is still far less than is justified by its importance in cold-regions hydrology. It is symptomatic that there is more emphasis placed on the subtleties of sediment-data collection, data that has somewhat limited application and no direct influence on the flow, than on measurements of the ice effects, which have a direct and important influence. However, as indicated by the above brief review, in recent years the pace of research on floating-ice hydrology has quickened and currently there is considerable activity.

5.4.1. Ice Jams, Break-Up and Freeze-Up

On the subject of ice jams there are now several long-term programs of observations of the processes involved in ice-jam behaviour. The longest program of systematic field observations of ice jams and break-up in Canada are likely those of the Alberta Research Council (ARC), begun in 1974 and described earlier. Of the various sites involved in this program over the years, the emphasis has been on the Athabasca River at Fort McMurray. This reach is the site of dynamic break-up and frequent ice-jam formation, and has immediate and practical importance due to the major threat ice jams pose for the City of Fort McMurray. The initial emphasis of these observations was on confirming the applicability of the equilibrium analysis to ice jams in such irregular field situations and on deducing the properties of the pack (Andres and Doyle, 1984). However, over the years the emphasis has evolved to an investigation of the break-up ice regime of the reach in general, initially with a view to developing a break-up forecast procedure for Fort McMurray (Andres, 1988), but with the longer-term view of understanding the break-up process in general. An example of the progression in development of open water upstream of Fort McMurray is shown in Figure 5.16. This is an important precursor of break-up.

NHRI is continuing a major study of break-up and ice-jam formation on the Liard River upstream of its Mackenzie confluence in the Northwest Territories (Prowse, 1986), a reach selected for its consistent ice-jam formation and dynamic break-up that is of major importance due to the role it plays in controlling break-up over the whole length of the Mackenzie River downstream of the confluence. The observations were initiated in the early 1980s and have been of gradually increasing sophistication. A unique and major emphasis of the, at times, quite daring observations has been detailed documentation of the energy budget of the ice - both the in-place ice sheet and the ice-jam pack. As described earlier, it is this program that has made the major contribution to understanding heat transfer to the pack by the warm water from upstream and its influence on pack melt, an important facet of ice-jam behaviour. The logistics of access and instrumentation along this reach have been developed to the point where break-up progression over a long reach can be followed in considerable detail, the objective being to develop a better understanding of the dynamic break-up process. The process of ice decay and melt play a pivotal role in break-up and, as a supplement to the observations and energy-balance studies on the Liard River ice itself, a program has recently been begun to document and quantify the process of ice decay and its effect on ice strength (Prowse *et al.*, 1988).

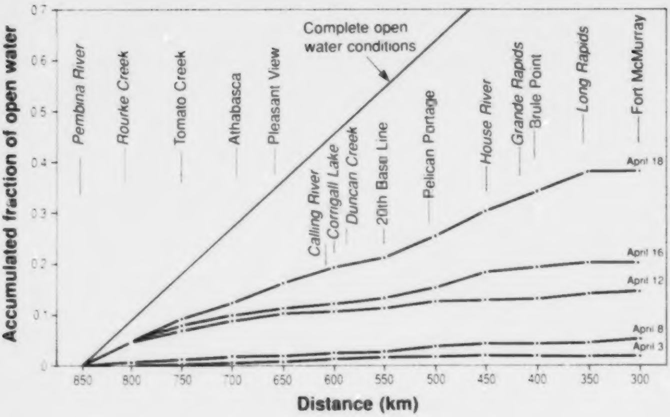

Figure 5.16. Development of open water prior to break-up over the reach of the Athabasca River upstream of Ft. McMurray, April, 1986 *(from Andres, 1988).*

The results of this latter study will have major significance in other fields as well as in the hydrology of floating ice.

Another program of observation of ice regime and related hydrological processes is that by the University of Alberta on the Hay River in the Northwest Territories (Gerard and Stanley, 1988b). This programme began because of the flood threat to the Town of Hay River, which is situated where Hay River enters Great Slave Lake. Hay River is a north-flowing stream of moderate size with a large variety of slopes over a short length and, for a northern stream, has good hydrometeorological information and reasonably good access for observations over a significant portion of its length. An ice-jam flood forecast procedure has been developed for the Town of Hay River and work is continuing on refining the procedure to allow for the effect of surges released by ice-jam failure upstream.

Systematic field observations of ice jams and break-up have also been carried out on the Thames River in Ontario (Beltaos, 1988). The Thames is a small low-gradient alluvial stream that runs through agricultural land. Ice jams can be expected to form each winter, especially in the reach just upstream of its confluence with Lake Huron. However, unlike the two northern reaches described above, this river is in a developed area, is small and is not entrenched. Consequently access along this river is not a problem, allowing detailed documentation of the ice processes. Because this is in a more temperate region, it is not uncommon for break-up to be initiated by mid-winter rain. The ice jams that form can subsequently freeze in place, allowing access for detailed documentation of the pack. This has recently provided the opportunity for the first thorough investigation and analysis of an ice-jam pack.

Other reaches in a temperate area that are the subject of detailed documentation of ice jams and the break-up process are the White and Connecticut Rivers near Hanover, New Hampshire which are being investigated by CRREL (Ferrick *et al.*, 1986). Unlike the Thames River, the White River is steep, with somewhat irregular bedrock controls, a characteristic of many rivers in New England and eastern Canada but, like the Thames, it is small and reasonably accessible, and can also have mid-winter ice runs and ice jams that form and refreeze. The special feature of the Connecticut River is that it is a medium-sized, readily-accessible stream extensively developed for hydro-electric power production. Its flow is almost fully controlled, with dams at frequent intervals along its length. This affords the opportunity for planned discharge releases to trigger break-up in very controlled circumstances, allowing detailed observation and documentation of the processes. This is a unique and valuable circumstance that has only recently begun to be fully exploited (Ferrick *et al.*, 1988).

The primary emphasis of the above studies has been break-up. The studies, being carried out on the St. Lawrence River near Massena by Clarkson University, are primarily concerned with freeze-up because of its importance in the operation of the large hydro-electric plants along this river. These studies have formed the basis of the detailed ice-regime program recently developed by the university and will doubtless continue to be used for its evaluation and refinement. A recent development has been a co-operative investigation by Clarkson University with an institute in China on a very detailed and personnel-intensive investigation of freeze-up processes on the Yellow River near Heque (Sun and Shen, 1988).

Another detailed investigation of freeze-up processes on a river in a temperate region is that of the Ohio, Illinois and upper Mississippi Rivers carried out as part of the River Ice Management (RIM) Program of CRREL (U.S. Army, 1990). These rivers form a major component of the extensive inland navigation system of the U.S.A. and shipping operations on these rivers are often impeded by ice conditions, both in the operations of the tows and of the many locks and dams. The initial focus of the study has been on the Ohio River. The most significant portion of this large programme as far as floating-ice hydrology is concerned has been the development, and field evaluation and calibration, of an algorithm to predict water temperatures and initial ice formation on this large river system which is characterized by a succession of pools large enough to limit mixing in the depth and complicate the temperature prediction (Daly, 1988).

Over the years many *ad hoc* studies of ice regime by consultants and others have resulted in a useful body of information. An example of the type of information often collected is given in Figure 5.17. Unfortunately the details are often left in the grey literature of reports to clients. Information is then difficult for others to obtain. However, a few efforts have been made to overcome this problem (Petryk, 1985, 1990; SEBJ, 1984).

In many cases the above field studies are being supplemented by laboratory and analytical studies. A particularly valuable contribution is the development of a non-proprietory synthetic ice at NWRI (Wong *et al.*, 1988). This should permit detailed investigations of the mechanics of break-up in the controlled environment of the laboratory, without the distraction of cold

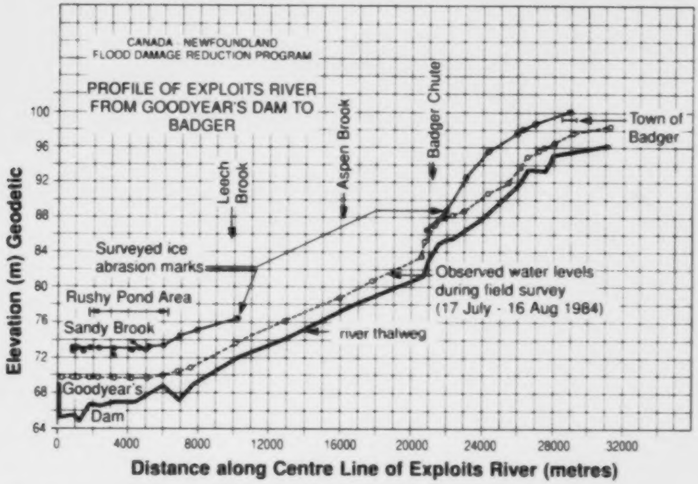

Figure 5.17. Freeze-up pack on the Exploits River near Badger, Newfoundland *(from FENCO, 1985).*

room temperatures and undesirable thermal scale effects. Wuebben (1988) has begun investigation in the laboratory of the interaction of an ice jam with an erodible bed, a problem of major concern with respect to possible scour in the vicinity of the ice jam (Mercer and Cooper, 1977).

Analytical and laboratory investigations are continuing at Clarkson University on the processes of lodgement at freeze-up (Tsai *et al.*, 1988). An understanding of this process is vital to prediction of freeze-up on a river. The University of Iowa also has a major laboratory program investigating the interaction of ice jams and bends (Aguirre, 1988).

Like lodgement, a major outstanding problem is the exact mechanism controlling dynamic break-up. The observations of this process on the Connecticut River by CRREL are being complemented by a major numerical study of the unsteady flow that is characteristic of break-up and its interaction with the ice cover (Ferrick *et al.*, 1986). A study with similar objectives is underway at the University of Alberta (Gerard and Flato, 1988; Williamson, 1989). While these break-up studies are primarily concerned with the driving mechanism for break-up, the studies of ice decay by NHRI are focused on the resistance to break-up. In addition to the field observations of ice decay and melt described earlier, NHRI have initiated a detailed laboratory investigation of the problem (Prowse *et al.*, 1990).

5.4.2. Low Flows

Considering the environmental significance of low flows and their importance in pollution studies, the effort being devoted to the understanding them in ice-affected rivers and lakes is in marked contrast to that of high stages associated with ice jams: there seem to be only a few active research programs concerned with winter stream conditions in North America (see, for example, Calkins and Brockett, 1988; Blachut, 1988).

5.4.3. Flow Duration

Efforts are continuing on the development of non-proprietary numerical models to allow continuous simulation of river- and lake-ice regimes. A major effort in this regard is the study currently being funded by the Canadian Government (Petryk, 1988). Other examples are the programs under development by Clarkson University, with funding from CRREL, and those in Finland and France mentioned earlier. Presumably the Swedish Power Company is also working on a similar model to complement their major field installations (Billfalk and Dahlback, 1985). A less ambitious, but nevertheless quite successful, example of development of a program to simulate freeze-up progression on a river, which attempts to capture the essence of the various processes while keeping the algorithm as simple as possible, is that by Andres and Spitzer (1990). There are doubtless other such programs available or currently under development. It is a very active field of investigation and when documented, calibrated, and made readily available these programs will represent a major advance in the field of floating-ice hydrology and, indeed, a very belated and important advance in the field of hydrology in general.

5.4.4. Flow Measurement

As in the past, a necessary component of the research on the above aspects of floating ice is accurate measurement of discharge and water level in the presence of ice. A valuable and ingenious contribution to this problem in a difficult environment are the continuing spring flow-measurement camps being held, as funds permit, by the Water Survey of Canada in the Northwest Territories as described earlier. Photographic techniques such as those described by Prowse *et al*. (1987), and the use of videos, will likely become more common as the significance of the freeze-up and break-up period is appreciated and equipment becomes cheaper and more reliable.

A vital need has been a more reliable method of measuring water levels during the dynamic ice conditions of freeze-up and, particularly, break-up. The standard bubbler-gauge installation is notorious for being rendered inoperative by damage to the bubbler line or movement of the orifice during an ice run. Instruments to overcome this by measuring water levels from above, rather than from below, have finally been developed (Andres and Cardinal, 1989; WSC, personal communication). These use an electronic range finder and are currently undergoing field trials. A typical installation is shown in Figure 5.18.

5.5. FUTURE RESEARCH NEEDS

There is no doubt the major stumbling block to an improved understanding of floating-ice hydrology is the current lack of an appreciation of the importance, and indeed the increasing importance, of the subject among hydrologists and others involved with hydrology in Canada. An apprecia-

Figure 5.18.
Water Survey of Canada field trials of downward-looking sonar for measuring water levels, Hay River at Hay River, 1989.

tion of this, and the other aspects of cold-regions hydrology discussed in this publication, will likely not come about until more emphasis is placed on Canadian hydrology in the hydrology courses taught in Canadian universities and colleges.

But in the meantime there is much more research needed. The following suggestions focus on what are believed to be bottlenecks in the subject; the list is not exhaustive.

5.5.1. Ice Jams, Break-Up and Freeze-Up

A crucial aspect of the simulations of ice regime that are so important for the assessment of flooding, planning of hydro-electric operations and simulations of winter flows in general, is an understanding of the nature of the development of an ice pack or, in more common parlance, an ice jam. Despite the important advances that have been made over the last quarter century or so, there are still important aspects of ice jams that are poorly understood. Some advances in understanding should be relatively easily accomplished in a regular laboratory setting, without the need of cold-room facilities. These would include: a better understanding of the toe region of an ice jam, including the magnitude and nature of the force applied by the ice jam to a restraining ice cover and the possibility and mode of 'grounding' in a shallow stream; the nature of the pack formed in a reach of strongly varying width and/or curvature; the strengthening effect on a pack of the refreezing that commonly takes place at freeze-up and, sometimes, at break-up; the nature of the 'entrainment', or stability, mechanism at the head of a pack (the current wisdom is that this is reasonably well-defined, but in fact relations currently used - usually a critical value of a form of 'pack densimetric Froude number' - have not been extensively evaluated and the nature of the interaction of the pack with an erodible bed, this latter being an important question with regard to pipeline crossings and bridge foundations.

However, by far the most important question with regard to break-up ice jams is when and where they will form in a given reach and when and how they will fail. It is a complex question, involving as it does an interaction between hydraulics, mechanics and heat transfer over a long reach. Its solution will require development of an understanding of the break-up process in general. As mentioned, an impressive start has been made on this but there is yet some way to go. The question can be reduced to four components: what does it take to trigger an ice run; what is the process by which an ice run is sustained, sometimes over hundreds of kilometres; under what circumstances will the ice run stall, so forming the ice jam that is of such concern; and, finally, under what circumstances will this ice jam fail and initiate another ice run? Further advances on these problems will take a continuation of the combination of mathematical modelling and programs of well-planned, long-term, reach-scale, comprehensive, dedicated and astute field observations in a variety of climatic and geomorphic circumstances. Progress can also be made by decoupling the thermal aspects and considering just the hydraulic and mechanical interaction, processes that are amenable to standard physical model studies in a laboratory, either using ice covers grown in a cold room with their strength scaled down by doping (Predmore, 1986), or using synthetic ice such as that developed by NWRI. The thermal aspects would seem to be best approached on the basis of numerical modelling and selected field and laboratory studies of the important processes involved, such as those being carried out by NHRI and ARC. But progress on this front will likely be slow because of the complexities involved. In the meantime more application of a straight empirical approach for specific reaches, based on a general understanding of the ice processes involved, historical hydrometeorological records and limited detailed observations of the ice processes in the reach, can prove worthwhile.

Break-up, ice runs and ice jams have begun to yield their secrets under such programs and, while the effort required for further progress is substantial, the reward will be a major advance in cold-region hydrology.

Questions of similar significance remain with regard to freeze-up. Like break-up, significant advances will require a combination of mathematical modelling, laboratory studies and well-planned field observations that include consideration of the mechanics, hydraulics and thermal aspects of the problem.

The major question at freeze-up is that of lodgement. It is the key to the successful simulation of ice-cover development, but as yet there has not been the concerted effort to solve it that includes the required detailed field observations.

Another important feature of a pack developing at freeze-up is the stability added to it by any solid-ice formation within the pack as the pack advances. As mentioned earlier, this has a major influence on the thickness of the pack that forms and thereby on the rate at which the pack progresses upriver. It can therefore have a large influence on the flow and water levels. While some progress on this subject can be made in a laboratory setting, an extensive program of field observations is also required.

Somewhat less pressing - but only because at least crude estimates of the processes are currently possible - is research on the production, transport and deposition of frazil ice, and the growth of border ice. While reasonable bounds can be placed on the production of frazil, very much less is known of its behaviour in the flow - particularly the effects of its active and passive stages - and its final destination in the waterway system. Similar comments apply to border ice. These questions have important implications when trying to model the development of the freeze-up pack, and thereby on the simulation of its influence on the flow and water levels.

5.5.2. Low Flows

The hydrological feature closely tied to freeze-up is low flow. This can be very sensitive to the progression of the freeze-up pack and, in small streams, to the growth of ice through the winter. A major unknown is the effect on low flow of ice-front advance, especially in reaches such as bedrock-controlled channels that are characterized by many lakes, rapids and other shallow reaches. The growth of ice in and over the controls of lakes and pools has been little investigated (see, for example, Carmack *et al.*, 1987).

Another surprisingly neglected question is that of the interaction of frozen ground and streamflow. For instance, it is currently not understood just where and under what conditions lakes and streams will freeze to the bed and isolate that portion of the stream system from ground-water input, so reducing base-flow to the mainstream.

5.5.3. Continuous Simulation

The ultimate objective is to be able to simulate the development and demise of the ice cover on a river, and the associated variation in flow and water levels, throughout the winter. As indicated earlier, a major need in this regard is for a non-proprietary numerical model to be developed and to be maintained as a national facility by an organization such as the Federal Government, which has the resources and long-term view to keep the model current and to encourage, fund and carry out the necessary research to improve the model. An initial effort with this end in mind is underway in Canada but, given the current state-of-the-art, this can only be a first effort. Much remains to be done.

5.5.4. Flow Measurement

A fundamental requirement for the direct use and verification of such a model is accurate discharge information. Currently the accuracy of water-level forecasts under ice conditions such as ice jams is often limited by the accuracy of the discharge information rather than by the model. Recognition of the importance of ice-influenced flows, and the need to collect reliable winter and spring streamflow data, is long overdue, as is the necessary expenditure and research effort to improve the situation. It would seem further systematic documentation of the problems of winter discharge measurements, and techniques to overcome them, should prove worthwhile. This is particularly so in view of the increased significance of winter flows in engineering.

At present many of the techniques of winter-discharge measurements are simply an extension of open-water measurement techniques. It is arguable whether this is appropriate in all cases. It would seem worthwhile to investigate the application in winter of techniques such as dilution measurements that, while they may be of marginal worth in open-water conditions, may be far more successful in winter. It is also possible that more use of a hybrid approach between field measurements and calculations could improve the discharge information, whether these calculations are direct hydraulic calculations for the flow measurement reach which make more rational use of a winter rating curve, or hydrological/hydraulic models of the catchment.

An immediate need is for more reliable information on discharges at break-up for periods of less than a day. While it has become accepted to determine the peak instantaneous discharge for open-water events, the same attention has yet to be devoted to the peak discharge during break-up, particularly in or near reaches known to be prone to ice-jam formation. Improvement in this situation will require that more attention be paid to documenting ice conditions and high water marks during break-up, through more on-site observations and measurements as has been done for the spring camps, or perhaps time-lapse photography. On large rivers there may even be a role to be played by remote sensing. Instrumentation other than the standard bubbler gauge, such as downward-looking sonar and ultrasonic velocity-measurement installations, should see more application. Improvement in even the bubbler-gauge operation may be possible if more elaborate installations of the orifice line and orifice can be made, much as used to happen with standard well installations.

5.6. NEW RESEARCH APPROACHES

It is difficult to anticipate new research approaches: these have a tendency to be developed as the need arises. However, there are some situations where the better and more extensive use of present understanding and technology should contribute to better research approaches, if perhaps not new ones.

A first suggestion is not a new research approach at all. It is a plea that it be recognized that much in the way of research is done by private consultants who, unlike University and Government research agencies, have no incentive to make the results available to a wider audience. Consequently it generally becomes lost to the research community, or at least very difficult to procure. A requirement for submission of the results of the work for possible publication in a refereed professional journal could be made part of appropriate government contracts, and funds made available as part of the contract to at least partially support the preparation of the publication. This would be a service to both the public and the profession, and would subject the work to wider scrutiny.

In the field in the past there has been a tendency to look at important events, such as the formation of an ice jam at a particular site, in isolation.

However it has become clear that such events are simply part of the big picture of the ice-regime on a river, and it is important for ice-regime research to take account of this. To make a significant contribution the approach must be reach- and even catchment-wide, and involve consideration of the hydraulic, hydrological, mechanical and thermal aspects of the situation.

To permit this reach-wide approach, yet avoid some of the large manpower requirements, much greater utilization of remote sensing, in the widest sense of the word, is desirable. There should be more use of periodic aerial reconnaissance (using a fixed-wing aircraft if necessary - it is very much cheaper than a helicopter and almost as effective) and of video coverage. On larger rivers the option of satellite surveillance has yet to be fully utilized. The work under the RIM program by CRREL illustrates what can be achieved using these techniques (Gatto, 1988).

Recent advances in electronic instrumentation and telecommunications also do not seem to be used to the full extent possible. As far as is known there is nothing in North America to match the extensive instrumentation employed by the Swedish Power Board to document flow and ice interactions that result from hydro-electric operations. At another level the use of radar to determine the thickness of ice covers and ice jams has yet to be developed and utilized to its capabilities. Even underwater video is a possibility, particularly at freeze-up: it has been used successfully before. Simple improved data loggers for use in the field under cold conditions have been a long time coming, as has the downward-looking sonar. More use should now be made of them to facilitate remote reach-wide documentation of events. These should soon be supplemented by reliable remote-video or time-lapse photographic coverage to allow records obtained under ice conditions to be interpreted.

As has been illustrated on the Connecticut and Hudson Rivers in New England, and in a more limited way on some rivers in Ontario (Foulds, 1988), much can be accomplished by field observations under controlled discharge conditions. Given the potential financial benefits to hydro-electric operations from a better understanding of ice conditions, such experiments would seem to be easily justified. There may even be possibilities for similar operations, perhaps on a smaller scale, that use irrigation storages and canals which, in North America, remain dormant in the winter.

In the past laboratory studies have tended to concentrate on modelling the various ice situations in the form of a true physical model. While this has been successful in many ways, it is also limited by the presence of large scale-effects. There is a need for laboratory research that begins to look at the hydraulic, mechanical and thermal fundamentals, such as stress transfer in a floating pack, stiffening action of re-freezing of a pack, dynamic friction along the failure surface between pack and bank and so on, with far less concern with precise simulation of an actual ice jam or other such situation. The results of such fundamental studies, together with mathematical models, could likely overcome many of the limitations of the pure physical model approach. Indeed, many of the principles of ice jams seem to have been deduced from laboratory experiments on log jams. Other examples are some aspects of frazil behaviour deduced from the behaviour of other particle types in a warm environment (Park and Gerard, 1984), and the interaction of frazil pans by investigating the interaction of air-supported disks on a chute (Ackermann and Shen, 1983).

While development of a computer model based on current understanding of a process hardly rates as research, there is no question that research into the ice regime of inland waters has been severely hampered by the lack of such a model. It is hoped the current Canadian effort to develop a model comes to rapid fruition.

The same can be said for hydrological models. Traditionally these have concentrated on simulations under open-water conditions, and particularly times of extreme discharges. However, the need in ice-regime research is for a hydrological model that allows reasonably accurate simulation of discharges in the transitional times of freeze-up and break-up.

A start has recently been made in Canada on field, laboratory and analytical research into the processes that contribute to ice decay. This effort is much overdue and deserves support and extension. In like manner, there is a strong need for field documentation of the detailed characteristics of the different ice formations that can occur, such as frazil, frazil pans, anchor ice deposits, ice dams and hanging dams (see, for example Beltaos and Dean, 1981), as well as river- and lake-ice covers in general. Estimates of the nature of these formations have in the past essentially been a matter of judgement. There is surprisingly little hard data. For example, the writer knows of no published work on the morphology of frazil pans, particularly the disposition of the slush under the pan, yet this is one of the essential inputs required of a rational model of freeze-up. Knowledge of the characteristics of these features will make an important contribution to simulation of ice processes and thereby to the hydrology of floating ice.

Another important feature that has been overlooked in the past is the stream thermograph. Much more work is required on the collection of data, and development of simple models, to allow the simulation of water temperatures in whole stream systems particularly during the ice year. As far as ice is concerned, the thermal regime can be as important as the hydrological regime.

The lack of information on the discharge in rivers during critical times of the ice year is a major limitation to the application of the current understanding of ice processes and to developing a better one. It is essential that operating agencies be made aware of the significance of discharges during the ice-covered portion of the year and that resources be provided or reallocated so that this portion of the water year can be given the attention it deserves.

Probably the greatest problem and the greatest challenge is the scale and complexity, and in many cases the inaccessibility, of the systems that must be studied. However, with the judicious application of: remote sensing, aerial (and perhaps underwater) reconnaissance, radar, telecommunications and data transmission, and other developments in field techniques that are rapidly becoming available; astute laboratory and theoretical studies of the fundamentals; and the application of the enormous computational capabilities that are now available, the problem can be rendered manageable.

Given a subject so fundamental to the fabric of Canadian hydrology, the effort is essential, the challenge exciting and the rewards inestimable.

5.7. CONCLUSION

In Canada, ice conditions can be as responsible for the two hydrological extremes of high water and low flows as the catchment processes of traditional concern. For this and other reasons, floating ice hydrology is a vital component of northern hydrology. While this seems to have been recognized in the early years of research in engineering and the natural sciences in Canada, when the community was more isolated from the fads and fashions of the rest of the world and more dependent on its own resources, it seems to have waned since. Certainly it has not been afforded the effort devoted to the subject in other cold-region countries such as the U.S.S.R. However, over about the last quarter century this interest has returned and research into northern hydrology, and the hydrology of floating ice in particular, has begun to take its rightful place in Canadian hydrological research.

The early focus of hydrology was on floods, and therefore the major target of early research in floating-ice hydrology was ice jams and the high water levels they caused. In recent years concern with the environment and pollution problems has begun to require that attention also be paid to low flows and water levels. In cold regions these features are particularly sensitive to floating ice hydrology. Major investments in enormous hydroelectric plants in cold regions have brought further pressure to develop an understanding of floating-ice hydrology because of the sensitivity of these plants to the influence of ice on flow and water levels throughout the winter.

This strong interest in the various facets of floating-ice hydrology has brought with it a demand for research to provide the understanding that will allow development of models to simulate flows and water levels throughout the ice season. This review of the state-of-research on this topic briefly discusses past accomplishments, present activities and future research needs from a Canadian perspective.

5.8. ACKNOWLEDGEMENTS

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6

GLACIER HYDROLOGY

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6.1. INTRODUCTION

The glaciers and ice caps of Canada are spread in a vast arc across the north and west of the country. They represent about twice the volume of the Great Lakes system and the presence of such ice masses within catchments has profound influences on northern hydrological regimes, particularly those of mountain streams. The importance of glacier hydrology to the economy is greatest from the point of view of water supply in western Canada where the waters from the Rockies, the Coast Range and other western ranges are vital for the production of hydro-electric power, for irrigation, agriculture and other industrial and domestic uses. In times of climatic change, when the elements in the water balance may change in their relative importance and the demands on water supply are increasing, understanding of streamflow generation with a view to better prediction of water supply is of paramount importance to the economy. Glacier response may also provide clues to the history of global warming and a method to monitor its progress.

The purpose of this chapter is, firstly, to describe the glacial hydrological system; secondly to illustrate the system with reference to Canadian examples and research; thirdly to describe the history of glacier hydrology studies in Canada, with particular emphasis on the more important recent work; and, finally, to examine future research needs.

6.2. GLACIER HYDROLOGY: AN OVERVIEW

6.2.1. Run-Off

The glacierized mountainous areas of the world have many characteristics in common with regard to their hydrology. Perhaps the most significant of these is that, even in dry regions of the North, they usually receive far more precipitation than the surrounding lowlands and provide the bulk of the water to many major river systems. Much of the run-off is derived from the seasonal reservoir of snowpacks, but it is complemented and sometimes exceeded by the long-term reservoir effects of glaciers. Although the general hydrological response of proglacial regimes has been reviewed in Sections 1.4.1.3. and 3.2.1., and Figure 3.1, the precise response can vary according to annual precipitation receipts. For example, in years of heavy snowfall, glaciers tend to grow and to withhold water from the rivers because of insulation provided by the snowpacks; in contrast, years of light snowfall are characterized by rapid ice-melt, glacier shrinkage and augmentation of river flow.

6.2.2. Local Variability in Climatic Inputs

Mountainous areas are characterized by heterogeneity in precipitation and energy inputs which have great influence on the mass balance of the glacier systems and ultimately, local hydrological response. Orographic influences lead to generally increasing amounts of precipitation with increasing elevation. The positioning of slopes relative to prevailing winds and moisture sources can result in great variability in moisture receipts from one side of a mountain range to another. Redistribution of snow by wind action and by avalanche activity makes for even more local variability in ice accumulation and run-off generating capacity. Energy receipts, greatly influencing timing and rates of melt, exhibit enormous local variability. Radiation receipts are highly dependent on the angle and azimuth of sun and slope; radiation absorption is greatly dependent on albedo. Air temperature varies with elevation. Wind speed, with its effects on sensible and latent heat transfers, is locally highly variable. More detailed discussions of the various heat components as they relate to glacial hydrology studies in the North are provided in Section 8.2.3.

6.2.3. Regional Variability

The major glacier resources in Canada are found in, and sometimes spanning, a number of climatic regimes; some parts of systems are much closer to moisture sources than others and some glacierized mountain ranges within systems are protected from moisture-bearing winds by other ranges. The seasonal effect of sea ice on moisture supply can be an added complication in the high latitudes (see discussion of North Water in Section 2.1.2.1.).

6.2.4. Temporal Variability in Climatic Inputs

Some mountain areas receive precipitation evenly distributed throughout the year, albeit sometimes in the form of snow and sometimes as rain, while other areas typically receive most of their precipitation concentrated within a period of a few months. General precipitation patterns affecting the northern latitudes have been reviewed in Section 1.3.2.1. and can be compared to the location of the major glacierized areas of the North in Figures 1.5, 1.6, 1.7 and 1.11.

6.2.5. Glacier Outburst Floods

In many mountain areas, catastrophic floods result from the temporary ponding of water behind dams of glacier ice; the breaching or undermining of the dam can lead to a sudden and often disastrous release of the impounded lake. Such events can be several orders of magnitude greater than normal rain- or snowmelt-induced floods on the same river system. All these factors lend emphasis to the overriding characteristic of glacier hydrology, namely heterogeneity.

6.2.6. Distribution and Characteristics of Canadian Glaciers

The distribution of Canadian glaciers is illustrated in Figures 6.1 and 6.2. They are widespread and their characteristics reflect the variety of environments in which they are found. The high-arctic glaciers of the Queen Elizabeth Islands and Baffin Island (Figure 6.1) tend to be massive, often low-lying and in the form of ice caps with valley glacier outlets. The climatic regime of relatively-low precipitation inputs, intense winter cold and melting restricted to the short summer season, often results in these glaciers having large inclusions of ice below 0°C, their alimentation frequently dominated by the formation of superimposed ice. Ablation may be affected by calving into the sea or proglacial lakes and, hence, run-off production is low. Even in this region, however, there are significant differences between glaciers. The ice caps of southeast Ellesmere Island and Devon Island are more extensive and more dynamic than normal for the region, possibly because of the adjacent, permanently ice-free North Water polynya. The glaciers of the Western Cordillera (Figure 6.2) are, for the most part, found in higher and more rugged mountains than the high-arctic ice caps. Here, there are few true ice caps, but many valley glaciers have their upper parts linked together to form, in some instances extensive icefields. The maritime glaciers of the Coast Ranges tend to be larger, with a greater altitudinal range, and hence more dynamic than the continental Rocky Mountains' glaciers. This reflects great differences in precipitation input, in cloud cover, and in radiative heat balances during the summer melt season.

The massive valley glaciers of the Yukon, with the greatest elevation ranges of glaciers anywhere in Canada, form an intermediate group between the glaciers of southwest British Columbia and the High Arctic. Here many are surging glaciers and have associated glacier-dammed lakes leading to differences in run-off regimes.

6.3. PROCESS AND REGIONAL EXAMPLES

To illustrate the variety of glacier-hydrological environments, two examples are provided: a continental-mountain glacier in the Alberta Rockies and a high-arctic glacier on Axel Heiberg Island. Aspects of glacier hydrology are illustrated through these examples and by references within the sections on the history of research. Most discussion focuses on the continental-mountain glacier because of the large quantity of available scientific data and the author's personal research experience at this site. Before reviewing the specific examples, however, it is useful to consider a simple generic model of glacier meltwater production.

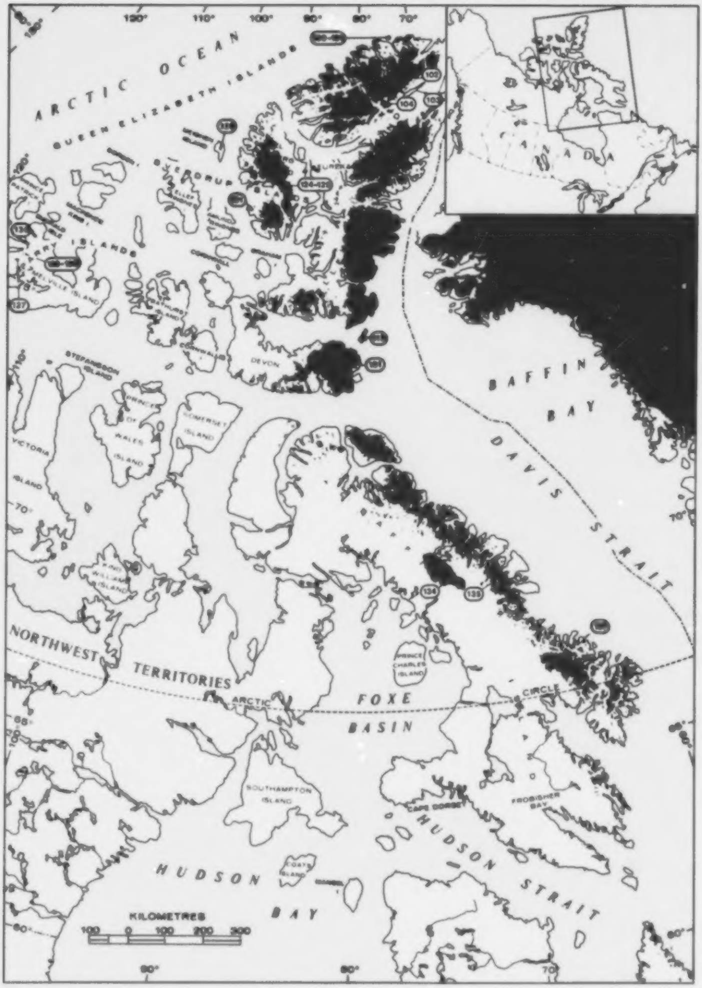

ARCTIC OCEAN
QUEEN ELIZABETH ISLANDS
SVERDRUP
PARRY ISLANDS
MELVILLE ISLAND
BATHURST
CORNWALLIS
DEVON
STEFANSSON ISLAND
PRINCE OF WALES ISLAND
SOMERSET ISLAND
VICTORIA ISLAND
KING WILLIAM ISLAND
NORTHWEST TERRITORIES
ARCTIC CIRCLE
FOXE BASIN
PRINCE CHARLES ISLAND
SOUTHAMPTON ISLAND
COATS ISLAND
CAPE DORSET
FROBISHER BAY
HUDSON STRAIT
HUDSON BAY
BAFFIN BAY
DAVIS STRAIT
CANADA
102
103
104
131
127
134
135
KILOMETRES
100 0 100 200 300
120°
110°
100°
90°
80°
70°
70°
66°
60°
60°
90°
80°
70°

Figure 6.1. Glacier distribution, Arctic Islands (see Table 6.4).

6.3.1. Typical Run-off Components from a Glacierized Catchment

Figure 6.3 shows a schematic annual discharge hydrograph of a stream draining a partially glacierized catchment. Discharge is composed of several components:

$$Q_t = Q_s + Q_p + Q_g + Q_i \qquad [6.1]$$

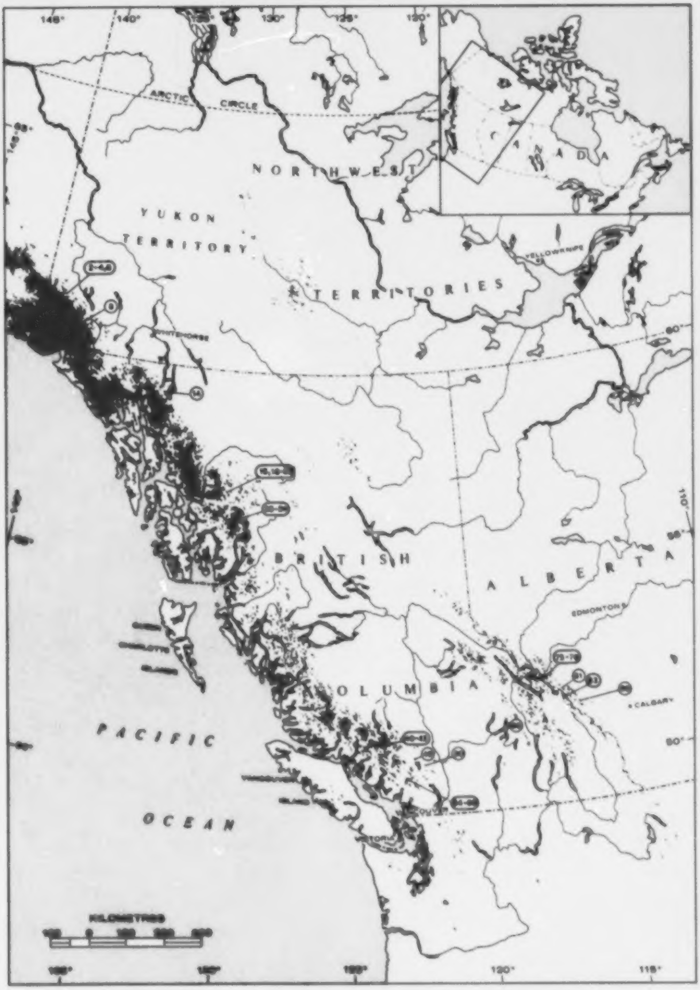

ARCTIC CIRCLE
NORTHWEST TERRITORIES
YUKON TERRITORY
CANADA
YELLOWKNIFE
WHITEHORSE
BRITISH COLUMBIA
ALBERTA
EDMONTON
CALGARY
PACIFIC OCEAN
QUEEN CHARLOTTE ISLANDS
VANCOUVER ISLAND
VICTORIA
KILOMETRES
140°
130°
125°
120°
60°
55°
50°
110°

Figure 6.2. Glacier distribution, western Canada (see Table 6.4).

where Q_t is total discharge, Q_s is snowmelt run-off from ice-free areas, Q_p is run-off from rainfall, Q_g is ground-water discharge, which may approach zero in highly-impervious catchments, Q_i is the glacial component of discharge which includes ice melt, firn melt and snowmelt from the ice-covered area.

The relative importance of each of these components, and the timing of melt, will vary according to the proportion of glacier cover within the basin

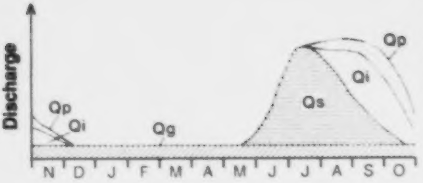
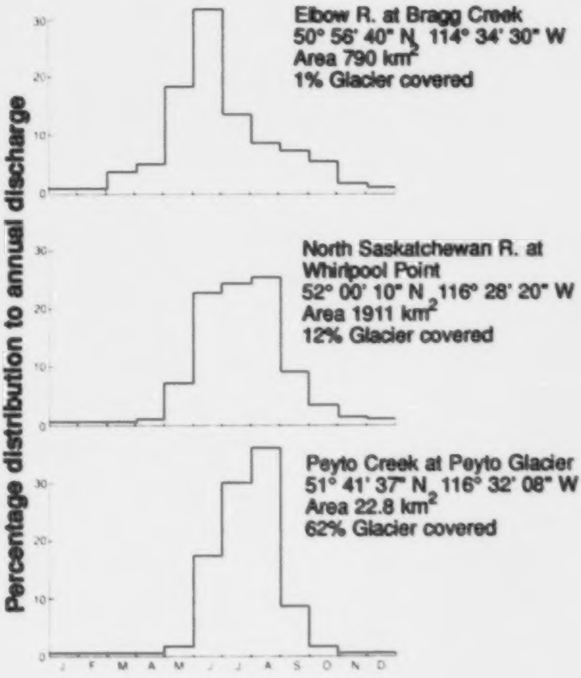


Figure 6.3. Schematic annual discharge hydrograph of a stream draining a partially glacierized catchment.



Elbow R. at Bragg Creek
50° 56' 40" N, 114° 34' 30" W
Area 790 km^2
1% Glacier covered

North Saskatchewan R. at
Whirlpool Point
52° 00' 10" N, 116° 28' 20" W
Area 1911 km^2
12% Glacier covered

Peyto Creek at Peyto Glacier
51° 41' 37" N, 116° 32' 08" W
Area 22.8 km^2
62% Glacier covered

Figure 6.4. Percentage contribution of mean monthly discharge to annual discharge, 1967 - 74, of three streams in the eastern Rocky Mountains.

and the general climatic setting: ice melt increases with increasing glacier cover (see Figure 6.4 with reference to Rocky Mountain catchments). In a continental situation (e.g., the Rocky Mountains and high latitudes) rainfall, snowmelt and ice melt are minimal in winter; in more maritime areas (e.g., the Coast Ranges) there may well be periods of rainfall and snow- and even ice-melt in winter. Further, in any particular catchment, one year may be very different from another. In general, years with heavy winter snowfalls and/or cool summers, result in minimal ice-melt while discharge derived from snowmelt will be relatively more significant. In years with light winter snowfall and/or warm summers, discharge derived from ice-melt will assume more significance. There is good evidence that conditions in spring and early summer are critical to explaining the hydrological response of a glacier. Cool periods in May and June, especially if accompanied by heavy snowfall, will maintain high albedo, absorption of solar radiation will be kept low, and ice melt will probably remain low all summer as relatively small areas of glacier become snow-free.

Figure 6.5 further illustrates the concept of the effect of proportion of glacier cover within a catchment on the hydrograph. For the summer of 1972 the hydrographs for the North Saskatchewan River drainage at Whirlpool Point (1911 km^2, 12% glacier-covered) and Peyto Creek at Peyto Glacier (62% glacier-covered) are shown. Peyto Creek is a sub-basin within the North Saskatchewan basin.

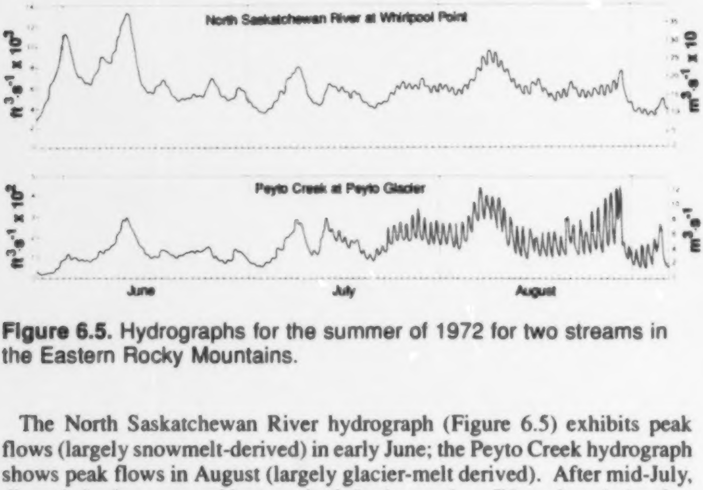


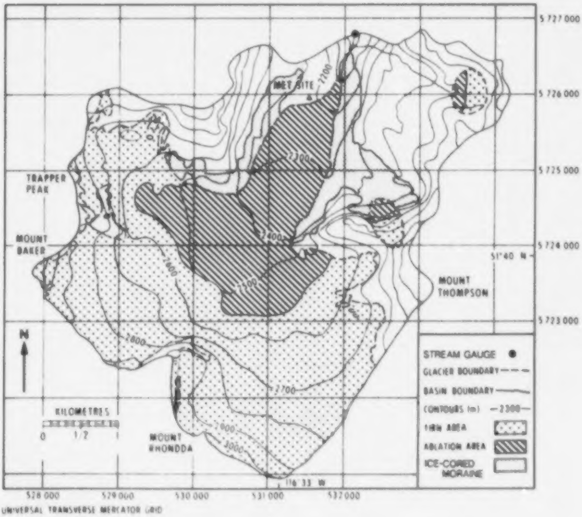
Figure 6.5. Hydrographs for the summer of 1972 for two streams in the Eastern Rocky Mountains.

The North Saskatchewan River hydrograph (Figure 6.5) exhibits peak flows (largely snowmelt-derived) in early June; the Peyto Creek hydrograph shows peak flows in August (largely glacier-melt derived). After mid-July, diurnal rhythms are evident on both hydrographs. Diurnal variation becomes more pronounced as the summer progresses especially on Peyto Creek. This is the result of an ever increasing area of bare glacier ice accompanied by widening of conduits within the ice through July and August leading to a very fast run-off response for the basin. In contrast, on the North Saskatchewan, important ground-water and non-glacier-derived flow components, in conjunction with the buffering effect of lakes on some of the major tributaries, result in a much more subdued diurnal oscillation.

6.3.2. Western Cordilleran Example - Peyto Glacier, Rocky Mountains, Alberta

6.3.3.1. Basin Description

The basin containing Peyto Glacier (51°41'N, 116°33'W; 22.8 km^2, 62% glacier-covered), lies on the eastern slope of the Rocky Mountains. Elevations range from the stream gauge, located at 1950 m a.s.l., to the highest peak at 3185 m a.s.l. The glacier and its environs have been studied extensively since the mid-1960s; analyses presented here pertain to the period 1967-74. Many of the concepts summarized in the following pages are from papers already published; specifically from Young and Stanley (1976b), Young (1981), Young (1982a), Munro and Young (1982). Ommanney (1987a) provides a bibliography of Peyto studies. The morphometry of the basin is shown in Figure 6.6.


5 727 000
5 726 000
5 725 000
5 724 000
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51°40' N
MET SITE
2200
2300
2400
2500
2600
2700
2800
2900
3000
TRAPPER PEAK
MOUNT BAKER
MOUNT THOMPSON
MOUNT RHONDDA
N
KILOMETRES
0
1/2
1
116°33' W
528 000
529 000
530 000
531 000
532 000
UNIVERSAL TRANSVERSE MERCATOR GRID
STREAM GAUGE ●
GLACIER BOUNDARY
BASIN BOUNDARY
CONTOURS (m) —2300—
FIRN AREA
ABLATION AREA
ICE-CORED MORAINE

Figure 6.6. Peyto Glacier Basin; elevation and terrain types.

The glacier terrain (the main glacier plus two small cirque glaciers within the catchment) may be divided into two zones - the ablation and accumulation areas. The ablation area is usually completely free of winter snow by late-July and is characterized by relatively dirty ice which develops well-defined conduits through which water exits the glacier very quickly. The accumulation area, characterized by firn and seasonal snowcover which lasts well into August, has much higher albedo and a much slower response time for water travel as meltwater is retained by the snowpack and released relatively slowly.

The non-glacier areas within the basin comprise areas of bare rock, ice-covered moraines and patches of semi-permanent snow. Each of these areas has its own particular snow-retention and snowmelt characteristics. The rock areas on the whole retain less winter snow than other areas, losing snow through blowing and avalanching. With much lower albedo and good drainage due to relatively higher surface slope angles, the rock areas also tend to lose their snowcover early in the summer.

Several of the sedimentary rock strata within the basin are permeable allowing a ground-water base flow throughout the year; lateral and terminal moraines vary considerably in their permeability.

6.3.3.2. Meteorological Inputs and Snow-Line Movement

Winter snow accumulation is of prime importance in determining quantity, timing and derivation of run-off during summer in a glacierized catchment. Heavy winter snowfalls usually lead to delayed run-off in spring due to high albedo and the retention of meltwater within the pack; the slow rising of the transient snow line leads to delayed melting of glacier ice. Quantities for net winter and annual balances by elevation band are given in Table 6.1.

Elevation (m a.s.l.)	Area (km²)	1967	1968	1969	1970	1971	1972	1973	1974	Mean 1967-74
31-3200	0.02	3.65	2.92	2.32	1.69	1.83	2.37	3.39	2.89	2.63
30-3100	0.17	3.26	2.60	2.10	1.54	1.73	2.21	2.94	2.54	2.37
29-3000	0.86	3.04	2.46	2.01	1.47	1.76	2.16	2.68	2.36	2.24
28-2900	1.77	2.73	2.20	1.82	1.35	1.61	2.00	2.38	2.13	2.03
27-2800	2.33	2.39	1.92	1.62	1.21	1.45	1.82	2.04	1.87	1.79
26-2700	2.62	2.12	1.69	1.47	1.09	1.36	1.72	1.80	1.69	1.62
25-2600	2.80	1.77	1.39	1.26	0.96	1.18	1.53	1.42	1.39	1.36
24-2500	1.25	1.48	1.15	1.08	0.84	1.03	1.38	1.16	1.19	1.16
23-2400	0.86	1.08	0.81	0.84	0.67	0.85	1.18	0.77	0.88	0.89
22-2300	0.65	0.88	0.64	0.13	0.59	0.75	1.09	0.61	0.76	0.76
21-2200	0.07	0.75	0.54	0.68	0.51	0.85	1.18	0.59	0.78	0.73
All glacier	13.40	2.06	1.63	1.43	1.07	1.31	1.67	1.72	1.62	1.56
Net specific winter balance by elevation band (m water equivalent)										

Elevation (m a.s.l.)	Area (km²)	1967	1968	1969	1970	1971	1972	1973	1974	Mean 1967-74
31-3200	0.02	3.65	2.92	2.32	1.39	1.83	2.37	3.39	2.89	2.60
30-3100	0.17	3.26	2.60	2.10	0.57	1.73	2.19	2.94	2.54	2.24
29-3000	0.86	3.04	2.46	1.89	-0.01	1.76	1.70	2.68	2.30	1.98
28-2900	1.77	2.14	2.14	1.16	-0.53	1.08	1.n9	2.38	1.81	1.41
27-2800	2.33	1.08	1.27	0.40	-1.09	0.36	0.44	1.60	1.05	0.64
26-2700	2.62	0.23	0.53	-0.28	-1.64	-0.32	-0.17	0.78	0.31	-0.06
25-2600	2.80	-0.89	-0.36	-1.06	-2.20	-1.02	-0.81	-0.39	-0.41	-0.89
24-2500	1.25	-1.77	-1.09	-1.73	-2.67	-1.64	-1.38	-1.24	-1.04	-1.57
23-2400	0.86	-3.03	-2.14	-2.67	-3.36	-2.52	-2.17	-2.52	-1.96	-2.55
22-2300	0.65	-3.63	-2.65	-3.14	-3.70	-2.98	-2.58	-3.08	-2.40	-3.02
21-2200	0.07	-4.01	-3.11	-3.60	-4.30	-3.64	-3.11	-3.33	-2.95	-3.51
All glacier	13.40	0.01	0.35	-0.40	-1.70	-0.41	-0.25	0.43	0.24	-0.22
Net specific annual balance by elevation band (m water equivalent)										

Table 6.1. Peyto Glacier mass balance 1967-1974.

Meteorological conditions during the melt season, in combination with the state of the basin at the end of winter, determine the rate and timing of melt as well as run-off due to rainfall. As shown by Röthlisberger and Lang (1987), net short-wave radiation is the primary control for melt rate on snow and ice surfaces. A complete radiation record is not available for Peyto, but mean air temperatures are given in Table 6.2 as a surrogate together with summary figures for precipitation. Of particular importance is that portion of precipitation which falls as snow.

		1967	1968	1969	1970	1971	1972	1973	1974	Mean 1967-74
Mean daily temperature at Base Camp (°C)	June	4.7	3.4	6.8	7.5	4.4	4.8	4.3	6.0	5.2
	July	7.8	7.8	6.3	9.1	8.0	6.3	7.9	7.7	7.6
	August	9.4	6.0	7.1	9.3	8.9	9.0	8.8	7.8	8.3
	June-Aug	7.3	5.7	6.7	8.6	7.1	6.7	7.0	7.2	7.0
Precip. at Base Camp (mm)	June	54.0	96.3	76.2	71.1	87.3	87.0	52.2	65.4	73.7
	July	54.9	42.3	75.6	78.3	45.3	26.4	55.5	34.2	51.6
	August	27.0	63.3	64.2	33.6	27.0	13.5	38.4	73.8	42.6
	June-Aug	135.9	201.9	216.0	183.0	159.6	126.9	146.1	173.4	167.8
Volume of Snowfall in Basin (v.e.) ($m^3 \times 10^3$)	June	1139	2972	2362	1388	2186	3237	1598	1793	2084
	July	1197	1157	1999	787	1264	318	777	489	998
	August	263	1055	1456	317	173	10	1038	695	626
	June-Aug	2599	5184	5817	2492	3623	3565	3413	2977	3709
Volume of Rainfall in Basin ($m^3 \times 10^3$)	June	1444	1533	1211	1942	1907	833	942	1277	1374
	July	1373	819	1536	2880	853	924	1829	1107	1415
	August	997	1909	1545	1256	1091	615	757	2764	1367
	June-Aug	3814	4261	4292	6078	3851	2372	3428	5148	4156
Volume of Precip. in Basin ($m^3 \times 10^3$)	June	2583	4505	3573	3330	4093	4070	2440	3070	3458
	July	2570	1976	3535	3667	2117	1242	2606	1595	2414
	August	1260	2964	3000	1573	1265	626	1795	3458	1993
	June-Aug	6413	9445	10109	8570	7474	5937	6841	8125	7864
Monthly Mean Snow-line Elev. (m a.s.l.)	June	2060	2110	2285	2285	2055	2065	2155	2060	2134
	July	2331	2348	2490	2595	2371	2334	2402	2320	2399
	August	2511	2506	2626	2802	2582	2511	2566	2494	2575

Table 6.2. Peyto Glacier monthly mean meteorological elements and snow-line movement, 1967-1974.

The timing of snowfalls as well as the vertical extent of the snow coverage is critical in early summer as this can delay upward movement of the transient snow line with important repercussions on albedo and hence on melt rates. It is important to recall that maximum possible daily short-wave radiation (under clear skies) drops off dramatically after the end of June; in mid-September, the maximum total available short-wave radiation is about 55% of that available on June 21. The importance of albedo in determining net short-wave radiation is made clear by comparison of Figures 6.7 and 6.8 where the position of the transient snow line is very apparent. A summary of the height of the transient snow line through time for 1967-74 is given in Figure 6.9.

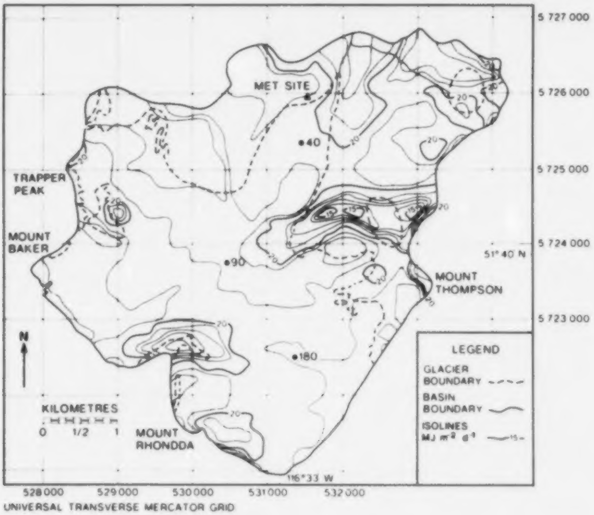
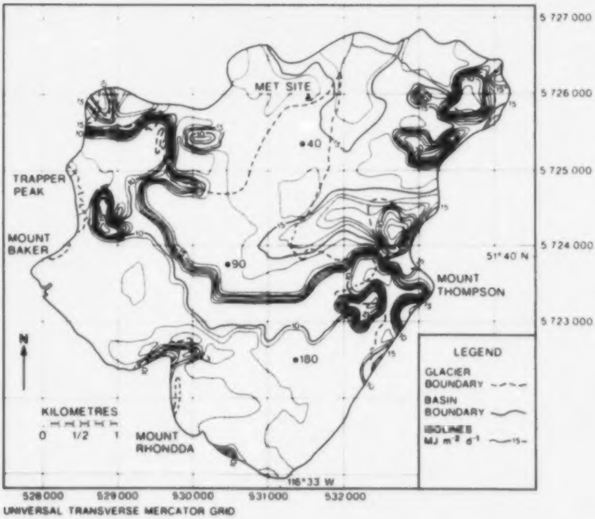

5 727 000
5 726 000
5 725 000
5 724 000
5 723 000
MET SITE
•40
•90
•180
TRAPPER PEAK
MOUNT BAKER
MOUNT THOMPSON
MOUNT RHONDDA
51° 40' N
116°33' W
N
KILOMETRES
0 1/2 1
LEGEND
GLACIER BOUNDARY
BASIN BOUNDARY
ISOLINES MJ m⁻² d⁻¹
20
15
528 000
529 000
530 000
531 000
532 000
UNIVERSAL TRANSVERSE MERCATOR GRID

Figure 6.7. Distribution of total global radiation on 1 August 1978, a partly cloudy day.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MET SITE
•40
•90
•180
TRAPPER PEAK
MOUNT BAKER
MOUNT THOMPSON
MOUNT RHONDDA
51° 40' N
116°33' W
N
KILOMETRES
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LEGEND
GLACIER BOUNDARY
BASIN BOUNDARY
ISOLINES MJ m⁻² d⁻¹
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UNIVERSAL TRANSVERSE MERCATOR GRID

Figure 6.8. Distribution of net short-wave radiation on 1 August 1978, with the transient snow line at about 2550 m a.s.l.

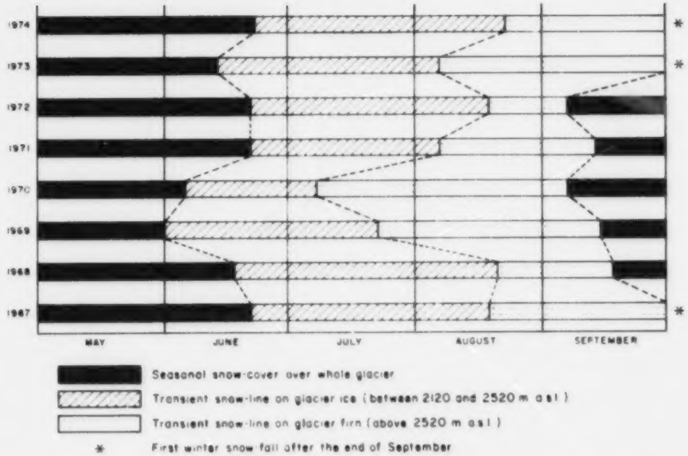

Figure 6.9. Variation in snow-line position, 1967-1974, Peyto Glacier.

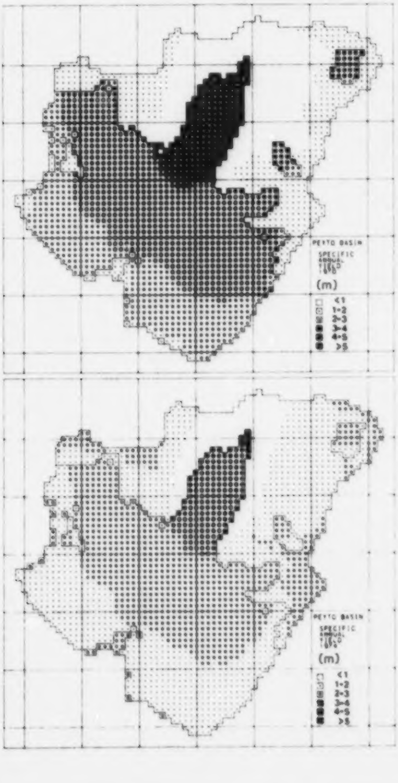
PEYTO BASIN
SPECIFIC ANNUAL YIELD 1970
(m)
<1
1-2
2-3
3-4
4-5
>5
PEYTO BASIN
SPECIFIC ANNUAL YIELD 1974
(m)
<1
1-2
2-3
3-4
4-5
>5

Figure 6.10. Maps of specific annual yield (m) for 1970 and 1974.

6.3.3.3. Internal Hydrology of the Glacier

During winter, melt is usually limited to that small amount generated by geothermal heat flow. Conduits and tunnels within and under the glacier close as a result of ice deformation. Residual water held within the firn and in crevasses and other cavities gradually drains out.

In springtime meltwater produced from snowmelt reaches the stream gauge only slowly, being retained in the snowpack and slowed in its passage through the glacier due to poor development of conduits. Tunnel enlargement is coincident with greater meltwater productivity (as the transient snow line retreats up-glacier uncovering glacier ice) and with less retention of water in the ever-thinning snowpacks. By late summer conduits are well developed resulting in fast transit times of meltwater to the gauge (Collins, 1982). This typically results in the timing of maximum instantaneous daily discharge moving forward from about 1700 hours to 1400 hours.

6.3.3.4. Specific Yield within the Basin

The net result of winter snow accumulation and meteorological conditions through the ablation season, in conjunction with movement of the transient snow line, is water yield which varies considerably within the basin. The two rather different years of 1970 and 1974 are taken as examples and specific annual yields within the basin are shown in Figure 6.10. The high yields from the glacier ablation area, especially in years of light winter snowfall, are apparent from this figure.

6.3.3.5. The Resultant Hydrographs

The hydrographs for 1967-1974 are given in Figure 6.11. Mean daily discharges in $m^3 \cdot s^{-1}$ are shown for total discharge, ice melt and firn melt. Volume of total run-off and volume of ice and firn-derived run-off are given in $m^3 \times 10^6$ for the months of June, July and August. The ablation season is divided into three parts corresponding to those in Figure 6.9:

1) when the whole glacier is snow-covered,
2) when the transient snow line is on glacier ice, and
3) when the transient snow line is above the firn line.

Table 6.3 summarizes the derivation of run-off for 1967-1974. It is apparent from these figures and tables that one year is very different from another for this small catchment.

6.3.3.6. Detailed Interrelationship for 1970 and 1974

Figure 6.12 summarizes, on a daily basis, the interrelationship between discharge and meteorological conditions for the two very different years 1970 and 1974. Discharge fluctuations closely follow temperature fluctuations (except in early-June 1970 when the melt from the snowpack was held within the pack). The effects of snowfalls during the ablation period, with a temporary drop in transient snow line, can result in reduced discharge for long periods of time. The heavy snowfalls in late-June and early-August 1974 are good examples of this effect.

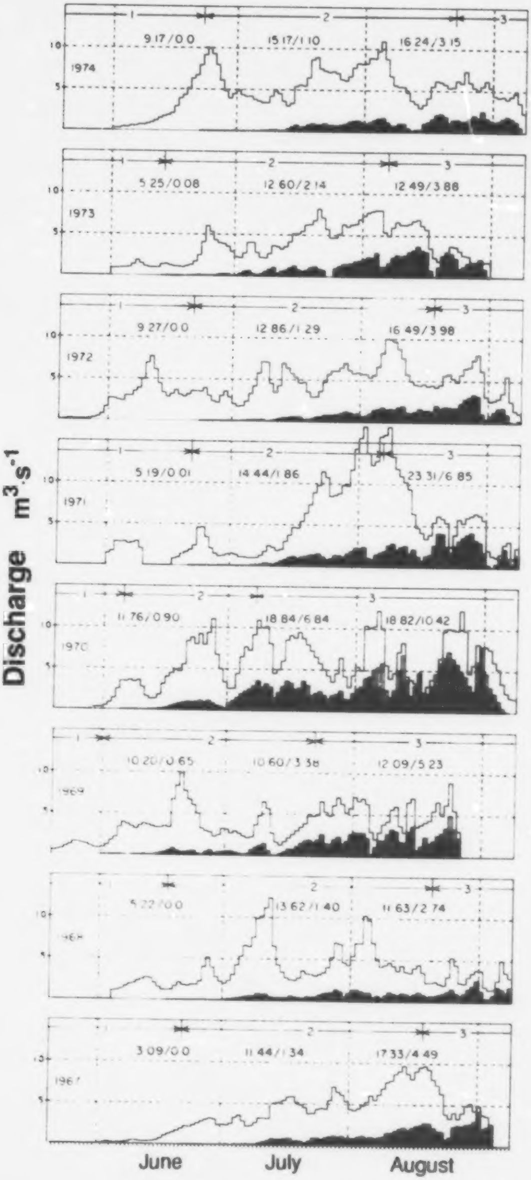

Figure 6.11. Hydrographs for Peyto Creek, 1967-1974 (Solid line indicates total discharge, striped area is the firn-melt component and black area the ice-melt component).

	1967	1968	1969	1970	1971	1972	1973	1974	Mean 1967-74
Glacier-derived run-off									
Ice melt (a)	14.5	10.9	14.9	18.8	15.4	9.5	13.5	11.5	13.6
Firn melt (b)	3.6	1.7	6.3	23.5	6.9	3.6	1.9	2.1	6.4
Snowmelt (c)	31.8	22.7	28.6	26.3	28.4	28.0	22.3	29.1	27.2
Summer precip. (d)	20.6	30.8	22.9	13.8	22.1	28.3	28.2	25.0	23.9
Total (e)=(a+b+c+d)	70.6	66.0	72.7	82.3	72.9	69.4	65.9	67.7	71.7
Non-glacier derived run-off									
Snowmelt (f)	16.2	14.3	12.4	8.6	12.6	11.8	16.1	16.2	13.4
Summer precip. (g)	13.3	19.7	14.9	9.0	14.5	18.8	18.0	16.2	15.5
Total (h)=(f+g)	29.4	34.0	27.3	17.7	27.1	30.6	34.1	32.3	28.9
Net Mass Bal. (10^6 m^3)	0.13	4.69	-5.36	-22.78	-5.49	-3.35	5.76	3.22	-2.90

Table 6.3. Peyto Glacier derivation of summer streamflow, 1967-1974: a - h as % of total calculated discharge.

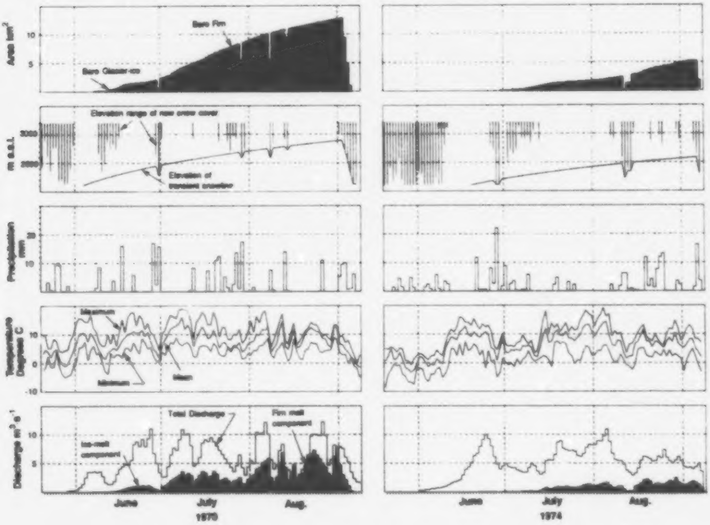

Figure 6.12. Snow-line movement, meteorology and discharge relationships, 1970 and 1974.

6.3.4. High-Arctic Example - White Glacier, Axel Heiberg Island, N.W.T.

White Glacier (79°27'N, 90°40'W; 37.95 km^2) ranges in elevation from about 50 to 1780 m a.s.l. It has been the site of a series of classic northern-latitude studies since 1959, most of the work deriving from the pioneering efforts of Fritz Müller.

Early work by Müller (1962) resulted in a vertical zonation model for accumulation and ablation in arctic glaciers, a model which, with appropriate modification, may be applied to glaciers outside the Arctic. Müller also developed methods for measuring ablation on glacier ice and, in a now classic study with Keeler (Müller and Keeler, 1969), showed how the accuracy of short-term ablation measurements can be greatly affected by changes from sunny to cloudy conditions.

An initial study by Adams (1966) on ablation and run-off on the White Glacier led to the later, more-detailed studies by Iken (Müller and Iken, 1973; Iken, 1977) in which meltwater production on both seasonal and diurnal scales is linked to variations in glacier velocity.

In separate, but related studies, Maag (1969) observed marginal ice-dammed lakes, two of which, Between and Ermine Lakes, are adjacent to White Glacier. Between Lake, dammed by both the White and Thompson Glaciers, was intensively studied during the 1960s and proved to be a marginal lake which emptied on a near annual basis, usually by overflowing the glacier and then cutting a meltwater channel quickly through the ice, but occasionally by exiting subglacially.

Other research germane to the colder high-latitude glaciers was the testing of the hypothesis that the internal glacier temperatures to a large extent determine the englacial hydrology. A comprehensive study by Blatter (1987) not only describes the internal glacier temperatures within the ablation zone, but shows that the ice close to the glacier margins and terminus is below 0°C to the bed, while the deeper ice in the centre of the glacier is not frozen to the bed. This paper, among others, has greatly increased our understanding of the linkages between englacial temperatures, glacier movement and the movement of meltwater within the glacier.

6.4. A HISTORY OF GLACIER HYDROLOGY STUDIES

The history of Canadian glacier studies is very short compared to studies in Europe, especially in the Alps, and even in comparison to the mountains of central Asia which have been inhabited and explored for many centuries. Glacier numbers in Figures 6.1 and 6.2 refer to Table 6.4 which lists glaciers on which significant measurement programs concerning mass balance and/or hydrology have been undertaken. An overview by geographical area is undertaken before a review of the effects of international programs sponsored by UNESCO (United Nations Educational, Scientific and Cultural Organisation).

6.4.1. Studies in the Western Cordillera

Simple glacier reconnaissance, including surveys of glacier snouts and photographic records, was initiated in a limited way at the start of the twentieth century in western Canada and was continued until the middle of the century. In the 1940s, the Water Survey of Canada started a systematic program of monitoring changes in the terminal positions of a number of glaciers in the Rocky Mountains; this eventually led to photogrammetric surveys of the ablation areas of a number of glaciers throughout the Cordillera. Volumetric changes in the lower parts of the glaciers could thereby be calculated (Reid, 1972). This program continued until 1980, but there

Glacier Name	Latitude	Longitude	Area km²	1940s	1950s	1960s	1970s	1980s
Yukon								
2 Hazard	61°15.7	140°21.9	16.0	....	....	....	....zz..z.	....
3 Steele	61°14.6	140°10.6	99.0	....	....	.....zzzz	zzzzz. z	z.........
4 Trapridge	61°13.6	140°20.0	5.5	....	....	......zzzz	zzz.z...	zzzzzzzzzz
6 Rusty	61°12.4	140°17.9	4.9	....	....	.....zoozz	zzzzzz....	....
8 Kaskawulsh	60°45.0	139° 6.0	700.0	....	....	zzzzzzzzzz	.zz.......	.....zzzz.
Coast Mountains (B.C.)								
14 Cathedral	59°20.3	134° 6.3	2.0	....	....	....	ooo.c.....	....
16 Alexander	57° 6.4	130°49.1	5.8	....	....	....	........oo	oooo....o
18 Yuri	56°58.0	130°42.2	3.6	....	....	....	.......ooo	oooo....o
19 Andrei	56°55.7	130°55.6	92.0	....	....	....	.......ooo	oooo....o
20 Forrest Kerr	56°54.1	130° 5.6	99.0	...xxxx.	....	....	.......ooo	o.........
* 23 Berendon	56°14.8	130° 5.0	33.4	....	....	....z.oooo	ooooo.....	....
24 Salmon	56° 8.6	130° 4.0	35.0	....	.......zz.z	....z.zzzz	....	....
40 Sykora	50°52.7	123°33.8	25.0	....	....	....	...ooooooo	....
41 Bridge	50°49.4	123°33.0	83.5	....	....	....	...ooooooo	....
42 Zavisha	50°48.4	123°25.3	6.5	....	....	....	...ooooooo	....
* 49 Place	50°25.3	122°36.0	3.8	....	....	.....ooooo	oooooooooo	oooooooooo
50 Wedgemount	50° 9.2	122°47.8	2.6	......x.x	x.........	...x.....x	.xx.xzxxxx	****xxxxx
54 Helm	49°57.8	123° 0.0	3.1	...xxxxxx.	x.x.x.x.x.		...ooooooo	oooooooooo
55 Sphinx	49°55.0	122°57.5	4.7	...xxxxxx.	x.x.x.x.x.	..x.x.x.x.x	...ooooooo	....
* 56 Sentinel	49°53.6	122°58.9	1.8	...xxxxxx.	x.x.x.x.x.	..xo*o*o*o	*ooooooooo	oooooooooo
Interior Ranges								
* 66 Woolsey	51° 7.5	118° 2.5	3.9	....	....	....oooooo	oooooo....	....
Rocky Mountains								
75 Saskatchewan	52°12.5	117° 8.2	30.0	...xxxxxx.	x.x.x.x.x.	xxxxxxxxxx	xxxxxxxxxx	....
76 Athabasca	52°11.7	117°15.0	15.0	...xxxxxx.	x.x.x.x.x.	xxxxxxxxxx	xxxxxxxxxx	....
* 81 Ram River	51°51.0	116°11.5	1.8	....	....	....oooooo	oooooo....	....
* 83 Peyto	51°40.6	116°32.8	13.4	.x.xxxxxx.	x.x.x.x.x.	x.o*oooooo	oooooooooo	oooooooooo
86 Drummond	51°35.5	116° 2.0	1.3	....	.x........	..****xx..	....	....
High Arctic								
100 Ward Hunt Ice Shelf	83° 7.2	73°30.3	660.0		...oz	.ozzzooooo	ooooo...o.	oo......
101 Ward Hunt Ice Rise	83° 7.0	74°10.0	32.0		........o	oooooooooo	ooooooo...	o.oo......
102 Gilman	82° 5.8	70°36.9	480.0		........o	oooooooooo	....	....
103 Unnamed	81°57.3	64°12.0	7.5			......oo.	oooooo....	....
*104 Per Ardua	81°31.0	76°27.0	4.3			...ooooooo	o.........	....
114 Agassiz	80°25.0	75° 0.0	17326.0	....	....	....	...sssssss	ssssssssss
119 Laika	75°53.0	79° 9.1	10.0	....	....	....	..ooo..oo.	....
*121 Baby	79°26.4	90°58.0	0.6	....	.........o	oooooooooo	oooooooooo	....
124 Thompson	79°28.0	90°30.0	230.0	....	x	xxxxxxxxxx	xxxxxxxxxx	....
*125 White	79°26.7	90°40.0	39.0	....	.........o	oooooooooo	oooooooooo	oooooooooo
126 Meighen Ice Cap	79°57.2	99° 8.0	90.0	....	.........o	oooooossoo	oo.ooooooo	oooooooooo
127 South Ice Cap	75°25.4	115° 1.1	66.0	....	....	..oooo.ooo	o.oo......	oooooooooo
128 West Ice Cap	75°37.8	114°45.0	36.4	....	....	..oooo.ooo	o.oo......	....
129 East Ice Cap	75°39.3	114°28.6	16.0	....	....	..oooo.ooo	o.oo......	....
130 Leopold	75°49.0	114°45.0	27.7	....	....	..oooo.ooo	o.oo......	....
131 Devon Ice Cap	75 20.0	82°30.0	12825.0	....	....	oooooooo.s	sssosooooo	oooooooo..
Low Arctic								
134 Barnes Ice Cap	70°10.0	73°30.0	6200.0	....	....	.zoooooooo	oooooooooo	.o........
*135 Decade	69°38.2	69°49.5	8.7	....	....	....oooooo	o.o.......	....
137 Boas	67°34.1	65°15.6	1.4	....	....	.........o	ooooo.....	....

* IHD glaciers

x = variations o = mass balance * = variations and mass balance z = other studies
s = other, some mass balance m = other, some variations

Table 6.4. Canadian glaciers having significant mass balance or hydrological records.

were shortcomings in this type of measurement. Snout retreat or advance may not be synchronous with overall loss or gain in total glacier volume. Indeed snout retreat may continue while there is build-up in the accumulation area, although in the case of a surging glacier, snout advance is normally accompanied by a volume loss in the accumulation area. Rate of snout retreat or advance may be more a function of local topography than of glacier nourishment (Haeberli *et al.*, 1989). Volumetric change as calculated from successive photogrammetric surveys represents net volumetric change for the area surveyed; the quantity of meltwater produced is always greater than this volume, the discrepancy being accounted for by glacier motion (although river discharge may be reduced by ground-water flow and evaporative losses). The early mass-balance work has been described by Østrem (1966).

6.4.2. The Arctic Islands

The history of investigations in the Arctic Islands is much shorter than the work in western Canada reflecting the much more recent opening up of the North. It was only in the 1950s that glaciological investigations of consequence were begun, involving long-term projects on some of the ice caps, (Devon, Meighen, Axel Heiberg, Ellesmere, and Baffin) and on some valley and cirque glaciers (e.g., Devon and Axel Heiberg Islands). Shorter-term studies have been undertaken on Melville and Coburg Island and on the Decade, Lewis and Boas Glaciers on Baffin Island; see for example Anonymous (1967) for work on the Lewis Glacier and Paterson (1969) for work on Meighen Island. A large body of literature developed around glacier/climate relationships in the Arctic; for example works by Andrews *et al.* (1972), Bradley and England (1978), Braithwaite (1981) and Keeler (1964).

6.4.3. Studies in the Yukon

Major glaciological studies in the Yukon have concerned glacier surges, climate studies (using ice cores) and floods from glacier-dammed lakes. The work on glacier motion and past climates has had only peripheral significance for the hydrologist; however, the massive movement of ice to lower elevations and the extensive break-up of the ice surface during a surge clearly influence meltwater yield to streams. Sub-glacial flow can also be redirected and, as in the case of Kaskawulsh Glacier, this can have important consequences for streamflow where flow may be directed from one major basin to another (Bryan, 1972). The long-term Icefield Ranges Research project incorporated many glacier-related studies. Recent glacial hydrology studies by Johnson (1985, 1986) and Johnson and David (1988) also deserve special mention. The considerable work on floods from glacier-dammed lakes by Clarke (1982) and Clarke and Mathews (1981) have made major contributions to our understanding of jøkulhlaups.

6.4.4. Studies in Labrador

There have been several geomorphology and mapping projects of glaciers in Labrador. In the recent past, glaciers have come under study from the point-of-view of mass balance and glacier/climate relationships (cf. Rogerson, 1986).

6.4.5. The International Hydrological Decade, 1965-1974

In the mid-1960s, under the auspices of UNESCO, International Hydrological Decade (IHD) studies were initiated which were to have a very marked influence on the conduct of glacier investigations (UNESCO, 1970). Glacier basins were considered a sub-set of hydrological basins and, because of the particular nature of the glacier environment, a special set of required measurements were devised. The major elements to be studied included the heat, ice and water balances within the basins as well as the interactions between climate and glacier response on short- and long-time scales. The following topics were singled out for special study:

a) Comparison of mass- and energy-transfer processes in snowfields and ice in various locations,

b) transformation and melting of snow and ice by thermal and mechanical means,

c) dynamics of glacier movement,

d) development of instrumentation for measuring volume and properties of snow and ice,

e) long-term mass balance of glaciers and ice caps,

f) erosion and sedimentation caused by glaciers,

g) discharge of snow, ice and water through a transverse section through a glacier, and

h) circulation of water in the interior of a glacier.

Recommendations were also made that, for intercomparisons on an international scale, three transects of research basins be set-up: one to circle the globe at about 45°N, another circumpolar chain in the Arctic, and a third north-south transect along the western side of the Americas.

The government of Canada, initially through the Geographical Branch and subsequently through the Inland Waters Branch, the Water Resources Branch and the National Hydrology Research Institute (NHRI), responded to UNESCO's recommendations by identifying six representative glacier basins: Sentinel, Place, Woolsey, Peyto, Ram River and Decade, at all of which a full set of hydrological, meteorological and glaciological measurements were made. These basins were chosen with a number of criteria in mind: they were to be typical, reasonably small, accessible, not heavily-crevassed, and preferably with some history of previous observation (Østrem, personal communication). The basins were also to be well-defined and have single-outlet streams which could be accurately gauged.

In addition, a number of research projects at other glaciers were set up and some of these, namely those at Berendon, Per Ardua, White and Baby Glaciers, were subsequently unofficially incorporated into the representative glacier-basin program. These have been described by Løken (1970) and Stanley (1975). The location, characteristics and a brief listing of major standardized data holdings are found in Tables 6.5, 6.6 and 6.7. For a more complete review of Canadian glacier-hydrology and mass-balance studies up to the early 1980s, reference is made to Young and Ommanney (1984) and Ommanney and Young (1988). To put Canadian studies in their worldwide context, reference is made to Collins (1984) and Young (1982b, 1985).

It is evident from the tables that there is considerable variability in the length (and quality) of the data records. Excellent records of basic heat, mass and water balances were obtained from Sentinel, Place and Peyto Glaciers; energy and mass-balance records from White and Baby Glaciers

Glacier	Major Region	Location	Latitude	Longitude	
Baby	Axel Heiberg Is	Expedition Area	79°26′N	90°58′W	
White	Axel Heiberg Is	Expedition Area	79°27′N	90°40′W	Arctic
Per Ardua	Ellesmere Is	Tanquary Fiord	81°31′N	76°27′W	Transect
Decade	Baffin Island	Inuqsuin Fiord	69°38′N	69°49′W	
Berendon	Coast Mountains	Stewart BC	56°15′N	130°05′W	North/South
Sentinel	Coast Mountains	Garibald Park	49°54′N	122°59′W	Transect in
Place	Coast Mountains	Birken BC	50°26′N	122°36′W	the Americas
Woolsey	Interior Ranges	Columbia Mts	51°07′N	118°03′W	West/East
Peyto	Rocky Mountains	Wapta Icefield	51°40′N	116°15′W	Mid-latitude
Ram River	Rocky Mountains	Front Range	51°51′N	116°11′W	Transect

Table 6.5. Locations of the Canadian glacier representative basins.

Glacier Name	Orient Ac	Orient Ab	Elevations (m) High	Elevations (m) Mean	Elevations (m) Low	Area km²	Length km
Baby	SW	SW	1175	1000	715	0.68	1.60
White	SE	SE	1782	900	53	37.95	15.30
Per Ardua	SW	SW	1705	1150	305	4.26	5.40
Decade	NW	NW	1467	1098	395	8.47	6.10
Berendon	E	E	2310	1350	650	33.40	10.00
Sentinel	N	NW	2100	1800	1525	1.99	2.20
Place	NW	NE	2610	2100	1820	3.96	4.50
Woolsey	NE	NE	2665	2238	1915	3.92	2.90
Peyto	NE	NE	3185	2635	2125	13.40	5.30
Ram River	NE	N	3020	2750	2560	1.80	2.10

Table 6.6. Physical characteristics of the glacier representative basins. (Ac = accumulation; Ab = ablation aspects)

Glacier Name	Basic Data Holdings Mass Balance	Basic Data Holdings Hydro-logical	Basic Data Holdings Meteoro-logical	Major Agencies Involved
Baby	1959-86			McGill, Trent, ETH Zürich
White	1959-86		1960-86	McGill, Trent, ETH Zürich
Per Ardua	1965-73			Dept. National Defence, NHRI
Decade	1965-73	1965-68		NHRI
Berendon	1967-69			NHRI
Sentinel	1965-88	1965-88	1965-88	NHRI
Place	1965-88	1965-88	1965-88	NHRI
Woolsey	1971-75			NHRI
Peyto	1966-88	1967-83	1967-83	NHRI, Ottawa, Manchester, Toronto
Ram River	1966-75	1970-74		NHRI

Table 6.7. Basic data collections and major agencies involved (McGill, Trent, ETH Zürich, Ottawa, Manchester, Toronto refer to universities).

were very good; the records at Ram River and Woolsey were shorter, but still good; records at Berendon, of heat and mass balance, were good for the two years of record (sufficient to accomplish the primary goal of estimating the likelihood of glacier advance); the records at Decade and Per Ardua were very short and have proven to be of minimal value for analysis. The most useful results came from the west-east transect in the southern Cordillera. Data summary reports were produced for Sentinel, Place, Peyto and Ram River Glaciers, Mokievsky-Zubok and Stanley (1976a, 1976b), Young and Stanley (1976a, 1976b); reports were not produced for Woolsey and Decade Glaciers. The Axel Heiberg Island studies are listed in Ommanney (1987). Results of mass-balance investigations were, however, reported for all Canadian glaciers through the Permanent Service on the Fluctuation of Glaciers, (Kasser, 1967, 1973; Müller, 1977; Haeberli, 1985; Haeberli and Müller, 1988) and have been summarized by Mokievsky-Zubok *et al.* (1985).

While basic data collecting was thus of variable quality, a perhaps unforeseen benefit from the program was that at some glaciers, particularly at Peyto and to a lesser extent at White, many additional special studies were undertaken on a variety of research topics. These studies greatly benefited from the provision of field laboratory facilities provided by the IHD program - indeed they might never have been conducted had not these facilities been available.

6.4.6. International Hydrological Programme (IHP) Glacier-Basin Studies

In 1975, the IHD program was extended into the IHP with important changes to program emphasis. The most significant of these was a realization that the research findings from glacier basins had to be integrated into mountain hydrology as a whole and that glacier/climate interactions would be more specifically emphasized and studied.

Studies on six of the ten glacier basins were terminated during, or at the end of, the IHD. On the remaining four, Sentinel, Place, Peyto and White, basic data collection continued with only little modification into the 1980s. The program on the West Coast was expanded to include mass-balance investigations on several other glaciers. In the Rockies, studies on Woolsey and Ram River Glaciers were terminated in 1975 so that more detailed studies could be conducted on the more-accessible Peyto Glacier and in its vicinity. Scientists from the Eidgenössisches Technisches Hochschule (ETH), in Zürich, and more recently from Trent and Queen's Universities, continued studies on Axel Heiberg Island, in particular on White Glacier.

It is mainly since the end of the. IHD that many of the detailed studies on glacier hydrological processes have been undertaken and written-up. Considerable understanding of glacier/climate interactions, internal glacier hydrology and discharge from glacier basins has been gained in this time. In addition, useful studies on glacier depth, ice movement and glacier mapping by remote-sensing and photogrammetric techniques have been completed. The data collected and the understanding of the processes involved have been and are being incorporated into streamflow simulation models such as the University of British Columbia watershed model (Quick and Pipes, 1977).

6.5. THE MAIN ACHIEVEMENTS OF HYDROLOGY STUDIES IN GLACIERIZED BASINS

6.5.1. Basic Description of Resources

In the last 20 years, there has been a substantial increase in basic data collection. The Water Survey of Canada, provincial and water-resource authorities, hydro companies and universities have considerably extended streamflow monitoring in mountain areas, increasing the length of record of long-standing gauges and establishing many new gauges on large and small mountain streams. Many snow courses have been monitored and, although most of these are in forested areas rather than the alpine zone, the information from them is very useful in streamflow modelling.

The Canadian Glacier Inventory proceeded steadily towards its aim of a detailed description of all glaciers until the program was shelved in the mid-1980s. Substantial sections within the Arctic and Cordillera have been described although there are still large areas left to describe and the data-storage system has yet to be fully integrated with other hydrological data bases. The inventory used maps and aerial photographs as prime sources of information; many sets of photos have become available in the past few decades (some of them commissioned for specific glacier projects) thus increasing the data base considerably.

Long- and short-term projects on specific glaciers have also done much to increase our basic descriptive knowledge. In the Cordillera, photogrammetric surveys of the lower parts of glaciers have, in some instances, been carried out frequently and in several cases systematically. Detailed information on mass balance, run-off and climatic inputs has been collected on several glaciers during the IHD and subsequently (on some glaciers similar data-collection programs existed independent of the IHD). Detailed, sometimes short-term projects, such as the North Water Project, the surveys on high-arctic ice caps, the projects on surging glaciers in the Yukon and on Berendon Glacier, have all added to the fund of basic data.

The sophistication and detail of meteorological data have increased substantially in recent decades as new monitoring stations were added to networks and as the advances in technology allowed more accurate measurements and faster and more sophisticated data dissemination and analysis. The combined balances projects all included meteorological measurements; these were taken both systematically at fixed sites over long time periods and, from time-to-time, at special sites as short-term experiments. While the consistency and quality of these data are not uniform, they nevertheless have been extremely valuable in explaining hydrological responses to meteorological conditions.

6.5.2. Achievements in Glacier Mass Balance

The objectives of the Combined Heat, Ice and Water Balances program of the IHD were outlined above and were partially successful. Several projects, for example that of White Glacier, Axel Heiberg Island, were incorporated into the combined balances project but did not meet all the basic criteria for inclusion. Results must be judged accordingly. The studies have really only been carried out during the last 20 years and many have been very short-lived. There are several points which should be noted in assessing the mass balance programs.

1. The glaciers studied have, for the most part, been easy to travel on with few very steep or crevassed sections and few areas prone to excessive avalanching. In this important respect the glaciers have not necessarily been typical of the surrounding glaciers.
2. Measurement sites on glaciers have usually been in areas of easy and safe access; steep head-wall areas, crevasse zones, areas of extensive wind-drift and avalanche zones have usually been avoided. This poses a major problem in extrapolation from measurement sites to unvisited parts of the glacier, especially on small glaciers of varied surface topography which are greatly influenced by local climatic effects. This is probably the greatest source of error in the overall balance figure for a glacier.
3. The accuracy of measurements varies considerably with climatic regime. Probably the most accurate mass-balance results are from temperate continental glaciers where the contrasts (in temperature) between winter and summer are greatest, where winter accumulation is relatively low and where assessment of mass balance at individual sampling locations is therefore relatively easy and accurate. It is very difficult on many of the west coast, maritime glaciers where winter snow accumulations can be extremely high (snow thicknesses of 10 m in places), definition of annual horizons in the accumulation areas of glaciers can be indistinct, and overall accuracy of balance estimates is often low. Complications in measurements due to superimposed ice and calving are often encountered on the glaciers of the High Arctic.
4. Some inaccuracies in the measurement program result from movement of marker stakes or, when there are no marker stakes, exactly the same measurement locations cannot be revisited each year; in some instances this has led to difficulties in comparing measurements made in different years. In many areas, especially in the lower parts of glaciers where changes in surface configuration can be large, this can be a major source of error, although no detailed studies have been done on the problem.
5. Often an approximate fixed-date system has been used for mass-balance measurements, however, this has not been the same fixed date for all glaciers thus complicating inter-glacier comparisons.

Despite these various provisos regarding mass-balance measurements, considerable knowledge and understanding has been gained. Understanding of glacier/climate interactions has substantially increased. Energy fluxes onto glacier surfaces (snow, firn, ice, debris cover) have been computed for many point surveys. The point-survey information has, in recent years, been extended to apply to entire glacier surfaces. Computer programs have been developed to simulate variation in radiation with topography (Munro and Young, 1982). Understanding of processes of snow drifting and deflation as a result of wind action have not advanced as far as our understanding of radiative heat fluxes and ability to simulate them. The mass balances of the continental glaciers have been shown to be greatly affected by energy inputs, the maritime glaciers responding much more to precipitation inputs. Trends in mass balance for the particular glaciers studied have been published for a number of glaciers and it is noteworthy that the occasional, very exceptional year can be of great significance for a glacier. While most years may be slightly positive or slightly negative in balance, a year in which low winter snowfall is followed by an unusually hot summer can show a strongly negative balance. Such years seem to be critical in the over-all reaction to climate and highlight the need for long series of observations.

In the early 1960s there was a notion that for a particular glacier the annual curves of specific net balance against elevation were parallel and that, once the slope of this curve was established for the glacier, measurements made at one location would be sufficient to define the curve for the whole glacier. It has been shown that for some glaciers this is not the case. While most years will exhibit curves of similar slope some years are exceptional and the curve is very different. This has considerable implications for mass-balance and run-off models.

Some mass-balance studies have been short, inconclusive and, in some respects, unsuccessful. The study on Berendon Glacier, for example, must be regarded as only a partial success as far as mass balance is concerned. Its heavy-snowfall maritime location underlined the great difficulties of mass-balance computations and, although new techniques (e.g., that of Mathews (1977) to relocate buried markers) were developed, the difficulties of assessing the mass balance accurately seem almost insurmountable.

R.M. Koerner, of the Geological Survey of Canada, and his colleagues, in their studies of the Devon and other arctic ice caps and glaciers (Fisher, 1979; Koerner, 1979, 1986; Koerner and Russell, 1979), have been interested primarily in glaciers as indicators of climatic change (both long-term from analysis of deep cores and contemporary from balance measurements at fixed locations). Their approach has been to develop elevation/balance curves based only on point measurements; inferences on balance quantities at unvisited parts of the glaciers are not made. While not making volumetric estimates for the whole ice mass, this method does give the best possible year-to-year comparisons; the method is simple and reliable and surveys can be made relatively quickly. The large ice caps, uncluttered by varied local topography, have the advantage that a measurement site is likely to be representative of much of the surrounding glacier surface; this contrasts with the small mountain glaciers where local relief and topography introduces the "noise" of local climates which must be filtered from the data record before major trends become apparent. Koerner's method could well be adopted more extensively in the future for those studies whose prime aim is to record climatic change through measurements on glaciers.

The complexities of glacier/climate response (both for mass-balance studies *per se* and for run-off production) have, to a considerable extent, been elucidated in the last 20 years (Young, 1981). The importance of the radiative heat flux, especially in continental locations, is well recognized as crucial to glacier "health". In this respect, albedo is all important:

movement of the seasonal snow line up-glacier during the ablation period largely determines the short-wave radiation flux and thus both mass balance and run-off production. The amount of winter accumulation and the frequency of summer snowfalls are critical in determining albedo.

The meteorological monitoring programs set up during IHD at sites on, or near to, glaciers have been only a partial success. Short-term, special projects relating energy fluxes to melt at sites have been successful (cf. Müller and Keeler, 1969; Holmgren, 1971a; Munro and Davies, 1977). Although great care has been taken to maintain consistently high standards of meteorological data collection, quality control has proven difficult. Moreover, most collection sites reflect the local climate at the site rather than average conditions on the glacier. Some studies have shown that as good as or better glacier/climate relationships can be established by using regular meteorological sites in nearby valleys than by using the data collected at the sites beside the glacier (Young, 1977). Letréguilly (1987) and Yarnal (1984a, 1984b) have conducted studies on glacier/climate relationships for glaciers in western Canada.

6.5.3. Achievements in Glacier Run-Off Studies

6.5.3.1. Meltwater Generation

Early investigations of Collier (1958) and Henoch (1971) in the Rockies took a regional approach to basin hydrology. These studies have led in the last two decades to more intensive studies on small glacier basins. Several of these studies have been designed to evaluate the relative contributions of radiative, latent and sensible heat transfers from the atmosphere to the glacier at point locations. Investigations have been made on snow (Föhn, 1973), glacier ice (Munro and Davies, 1977; Derikx, 1975) and debris-covered ice surfaces (Nakawo and Young, 1981, 1982). Most of these studies were undertaken for short periods during the ablation season and at only one or a limited number of sites on a glacier. More recently, as a result of advances in computer capabilities, the results at point locations have been applied over entire glacier surfaces and the effects of movement of the transient snow line on surface albedo have been incorporated into models. Such models will be useful in the future for transposing results between glaciers.

A great deal of useful ablation data has been collected at stake locations on glaciers. This information can be aggregated over periods long enough to overcome the well-known difficulties in interpreting short-term ablation measurements (Müller and Keeler, 1969), and can be correlated with standard meteorological measurements such as mean air temperatures. These results have proven useful for computer simulation models of run-off generation (Power and Young, 1979).

On both high-arctic and temperate glaciers successful studies of water movement in conduits within and under the glacier have been conducted (Iken, 1972, 1973; Müller and Iken, 1973; Clarke, 1986a; Shoemaker, 1987; Shoemaker and Leung, 1987). Dye-tracing and isotope methods have been particularly useful in accounting for discharge resulting from a variety of source areas (cf. Krouse, 1974). Understanding of water derivation (snow, firn, ice, rain) and the paths that water takes (englacial, sub-glacial, surface water, ground water) have increased considerably (cf. Collins, 1982; Collins and Young, 1979, 1981). Knowledge of processes at the glacier bed has developed remarkably as evidenced through conferences on the subject (see Paterson, 1986). The number of hydrographs now available for analysis in conjunction with meteorological, hydrological and glaciological data collected simultaneously has grown tremendously in recent years.

6.5.3.2. Glacier Outburst Floods

There are now several, well-documented examples of jøkulhlaups in Canada. These are from very different areas: the High Arctic, the Yukon and the Coast Ranges of British Columbia. Understanding of the phenomena has increased considerably with development of physically-based models. In particular, G.K.C. Clarke and his associates at the University of British Columbia, working on examples from the Yukon and British Columbia, have developed simulation models which are applicable world-wide. For an excellent overview of the history of outburst-flood investigation, the hydraulics of the phenomenon, the possibility of simulating floods, and reference to several well-studied Canadian examples, reference is made to Clarke (1986b). Other key Canadian contributions to outburst-flood studies by Canadians have been those of Ballantyne and McCann (1980), Blachut (1978), Blachut and McCann (1981), Clague and Mathews (1973), Clarke and Waldron (1984), Fisher (1973), Gilbert (1971), Maag (1969), Mathews (1965, 1973), Johnson and Power (1985) and McCann and Cogley (1977).

6.5.3.3. Models of Basin Run-Off

The advances in computer technology have facilitated the development of integrated basin run-off models. The University British Columbia VOLCAST and FLOCAST models (Quick and Pipes, 1977) are being used operationally along with models developed elsewhere. These models are partially physically-based but there are empirically-based relationships embedded in their structures. This reflects that there are still many unknown parameters in glacial and nival hydrology. Quantities of snow accumulation and rates of melt in glacierized zones remain difficult to determine; understanding of snowmelt processes in glacier firn areas is only partial; understanding of melt processes for glacier ice is now quite advanced but methods of transposing that knowledge to groups of glaciers remain to be devised. The work of Ford and his colleagues from McMaster University in the vicinity of the Columbia Icefields has shown the great importance of ground-water flow from one basin to another in some parts of the Rockies. This work highlights the need for care in interpreting results and in modelling procedures in areas of permeable bedrock.

6.5.4. Achievements in Surveying and Remote Sensing

This section might well be sub-titled "the effects of technological advances" for these have been very great in surveying and remote sensing, subjects greatly affected by the electronic revolution.

For the hydrologist interested in documenting changes in glacier volumes, probably the most exciting advances have been in the creation of digital terrain models of glacier surfaces by automated reduction of terrestrial or aerial photographs (Young and Arnold, 1977). Glacier ablation areas can be very accurately mapped using orthophoto techniques. Unfortunately, as with all photogrammetric methods, in the snow-covered areas of the glacier definition is poor and thus accuracy of mapping is low (Konecny, 1966). As a means of quickly and accurately mapping many ablation areas, and comparing with maps made at previous dates to arrive at volumetric changes, this technique holds much promise.

Mapping transient snow lines on glaciers from aerial and satellite surveys has also been undertaken (Østrem, 1973, 1975) and the paper by Howarth and Ommanney (1986) illustrates the considerable potential of remote-sensing techniques. Although the information so gathered has not yet been fully integrated into run-off models, there is considerable potential in these methods.

6.6. THE FUTURE

While many basic data reports have been published and papers written on a great variety of research topics, there are still many results to be written-up from the investigations already undertaken. To a considerable extent, there has been an overemphasis on data-collection activities without the necessary analytical follow-up. Although there have been several studies comparing glacial meteorological and mass-balance studies on a world-wide basis, there has yet to be produced a good comparative study of the extensive work carried out on Canadian glacier mass balance and glacier/climate relations.

In addition, there has been no real effort to evaluate just how "representative" the study basins are. The fact that the glaciers chosen had to be easy of access and easy to move around on, while neighbouring glaciers are commonly steeper and heavily-crevassed, begs the question of whether the combination of processes active in the research basins is in fact typical for the region. It can reasonably be argued that the same physical processes are active on all glaciers and thus the results of the studies undertaken can be applied to all glaciers. However, the precise way in which results are to be transposed to glaciers with substantially different surface topographies is still a largely unresolved question. This is a major prerequisite for the application of the results obtained from the small glacier basins to larger basins containing dozens or hundreds of glaciers.

The glacier studies must be integrated into the general context of high-mountain hydrology; this is a necessary prerequisite to the incorporation of IHD/IHP results into forecasting models relating to extensive regions. Considerable activity is taking place in this area of research and development. The World Meteorological Organisation, over the past several years, has undertaken intercomparisons of run-off models; this exercise has encouraged the integration of rainfall, snowmelt and glacier-ice melt into more general models.

While the results of these studies have as yet been little used by water managers (partly because of the lack of integration and partly because investigation of mountain hydrology is not seen as a high-priority activity), it is likely that these will be required in the future when climate change begins to affect the glaciological regimes. The applications of some of the latest General Circulation Models, to model a doubling of CO_2 over the next 50-100 years, suggest major changes in the water balance of glaciers in the western Cordillera and the North (see Section 1.6.2.1. and 1.6.3.). Baseline data collected at glaciers over the last two decades will prove invaluable for monitoring these predicted changes.

Much has been learned in the past few decades but much still needs to be done. While there have been considerable advances in some areas, there have been many problems in others and the limitations of measurement have been clearly demonstrated. There seems, however, to be considerable hope held out by new, computer-based technology, especially that linking remotely-sensed data with digital terrain models.

Our understanding of the interaction of meteorological variables on the glaciers to produce run-off has been greatly expanded by the combined balances projects. Much of the data from these projects has not been fully analysed or utilized; this analysis should be done and intercomparisons made between glaciers in conjunction with the collection of additional data. Glaciers preserve the climatic record and some studies, both on arctic and temperate glaciers should be maintained to add to the long-term records which will eventually prove of great value. The long-term maintenance of mass balance collection at specific glaciers, as advocated by Koerner (1986), seems particularly important. Some well-studied glaciers should also be maintained as field laboratories for detailed investigations on various aspects of glaciology.

Almost all mass-balance, hydrological and surveying studies in the accumulation areas of glaciers have proven difficult and frustrating. The inaccuracies of measurements in these areas, especially in high snow accumulation zones, should be clearly recognized and hopes of accuracies in measurement should not be raised falsely. Recognition, too, should be made of the difficulty of collecting good snow-accumulation and melt data from mountainous regions. The need to be able to predict seasonal streamflows for hydro-electric generation and other uses will grow as the demand for water rises. There will be a need, too, to have a better understanding of the long-term reservoir role of snowpacks and glaciers and their effects on stream nourishment. Projects to assess the impact of climatic changes on glacier streamflow generation should be undertaken.

Data banks of hydrographs, meteorological variables and glacier inventory information all exist. Understanding of many of the critical processes is considerable. New surveying techniques are available and are waiting to be applied. Run-off models have already been developed as a basis from which to proceed. The challenge is now to bring all this together to develop run-off models for large, integrated glacierized basins. ❄

7

WATER-QUALITY RESEARCH

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7.1. INTRODUCTION

The aquatic quality of Canada's northern waters is a scientific frontier. Historical activities, dating back to the 1940s, concentrated on describing the geochemistry and fisheries resources of these waters. However, as environmental concerns advanced the science of water quality within most parts of Canada, the aquatic systems of the North have to a large extent been relegated to infrequent, routine, chemical characterizations of the water resource. The reasons for the lack of activity, not necessarily a lack of interest, in the North are related to the following:

- a prevailing attitude that the sparse population and minimal industrial pressure will not significantly degrade water quality;
- the failure of many scientists and their supporting organizations to recognize the fundamental differences likely to occur in northern regions with respect to the fate and effects of local anthropogenic pollutants despite the obvious climatic and hydrological differences between northern and temperate regions;
- the high costs and logistical difficulties of conducting either monitoring activities, special studies or research programs in the North; and
- a failure until very recently to recognize the potential for environmental degradation in the North as a result of long-distance riverine, atmospheric and marine transport of pollutants.

As with other aspects of hydrology reviewed in this book, it is difficult to provide a cohesive and comprehensive review of the research activities in the Canadian North due primarily to the fragmentation of the work. The range of work considered has to be selected to provide a historical perspective, a spatial overview and a focus on key research activities. The geographical extent of this discussion will follow the 60°N definition described in Chapter 1. There are large biogeophysical differences throughout this massive region due to latitude, elevation, proximity to the moderating influence of oceans as well as to regional and local factors. This complexity will be minimized here in the interest of brevity and because of a general absence of information at this level of specificity.

Although this chapter concentrates on the Canadian North, international work is identified where relevant to the discussion. The content is restricted to the abiotic system, referring to the biotic system only as it may be directly related to aquatic chemistry. Consequently, about half of the "water quality" information are not presented here although the references include many of the relevant citations for the areas highlighted.

The survey of northern water quality will be presented in the following manner:

- a review of general research undertaken essentially prior to 1969 with the exception of research in basins or on subjects that will be covered in more detail;
- an overview of the water quality of northern river systems using Environment Canada's Northwest Territories monitoring network data set;
- a review of the Mackenzie River basin north of 60°N;
- a review of activities in the Yukon River Basin; and,
- a summary of two special research areas, specifically Resolute on Cornwallis Island and Saqvaqjuac on northwestern Hudson Bay.

Section 7.3 considers some of the more recent research activities related to local and long-range transport of trace organic compounds, hydrocarbons and acid deposition. The concluding section outlines future research and research support requirements.

7.2. HISTORICAL BACKGROUND

7.2.1. Activities prior to 1969

There have been a number of historical reviews of the water quality and limnology of arctic rivers and lakes. These have tended to depend on information from Alaska and the Scandinavian Arctic. One of the earliest was by Rawson (1953) who noted that "as yet there has been very little done in the Arctic". The report on "Industrial Waters of Canada" (Leverin, 1947) did not include surface water stations within the arctic region. Rawson (1953) further noted that much of the information for the Canadian Arctic had to be derived from "general accounts of surveys or explorations and from technical reports in various scientific fields". One example of this is the information contained in a series of articles by various authors in the 1947 Fisheries Research Board of Canada Bulletin entitled "North West Canadian Fisheries Surveys in 1944-1945" (e.g., Rawson, 1947).

Miller (1947) and two others surveyed the fisheries of Great Bear Lake between June 23 and September 1, 1945. During this time they covered over 2000 km (1250 mi.[a]) on the lake and although the major effort was directed toward the fisheries they recorded water temperatures, CO_2, O_2, pH and transparency. The relative simplicity of this survey is demonstrated by the use of a "small, white enamelled dinner plate suspended by a rope" for measuring transparency.

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[a] *The original documents recorded depths and distances in feet and miles: these, and the appropriate metric equivalents, are reproduced here.*

A similar survey conducted during the summers of 1944 and 1945 was reported by Rawson (1947) for Great Slave Lake. He subsequently reported additional, and apparently the most recent, whole-lake data on the physical and chemical limnology of the lake (Rawson, 1950). The characteristics of Great Slave Lake are more complex than those of Great Bear Lake as the former receives turbid water from the Slave River which only mixes in the western part of the lake to a distance of about 40 km (25 mi.[a]) from the delta. Thermal stratification in the main lake is generally observed from mid-July to late-August with the thermocline between 10 and 18 m in mid-August. Transparency is highly variable at less than 1 m in the vicinity of the Slave Delta to 4 to 13 m and 10 to 17 m in Christie Bay and McLeod Bay, respectively. Oxygen is abundant in the lake waters. The pH in the main lake is between 7.7 and 8.3 while McLeod Bay is lower at pH 6.6 to 6.9. The eastern parts of the lake, Christie and Mcleod Bays, receive drainage waters from the Precambrian Shield regions resulting in lower pH and very clear water with total dissolved solids (TDS) concentrations of 22 ppm as compared with 150 ppm in the main lake. The fisheries of the Yukon and Mackenzie Rivers were described by Wynne-Edwards (1947a, 1947b) in the same volume, but the surveys apparently did not include measurements of the water quality of these two river systems.

Other early studies of water chemistry in the North were undertaken by the Canadian Department of Mines and Resources and tended to focus on water geochemistry. Most of this work, as with the fisheries surveys, centred on the Mackenzie River basin. Thomas (1957) reported on the first survey of the waters of the Mackenzie and Yukon Rivers conducted during 1952 and 1953 (parameters included pH, colour, turbidity, suspended soils, dissolved solids, conductance, ions and alkalinity). Livingstone (1966) has summarized major ion data as well as more detailed information from Alaska and Greenland. Thomas (1964) presented for the first time the surface-water quality of the major drainage basins and northern areas of Canada from an industrial and developmental perspective. Single sample data were reported for the Yukon River at Dawson, the Mackenzie River at Aklavik, the Liard River at Fort Simpson and the Coppermine River at its mouth. These data were restricted to pH, colour, major ions, alkalinity and specific conductance. Of greater interest here was the first report of the chemistry of surface waters from a number of sites in the Arctic Archipelago. Based on the very limited available data, Thomas (1964) noted that the water was most commonly high in alkaline-earth bicarbonates. He also noted that colour may be high in tundra drainage from the mainland arctic coastal plains.

7.2.2. Overview of N.W.T. Water-Quality Network Data

Since the 1960s, the Water Quality (WQB) and Water Resources (WRB) Branches of Environment Canada have conducted a water-quality monitoring program within the N.W.T. The 20 stations presently comprising this network are located on Figure 7.1. The network was a response to the need for more thorough information on the water quality and quantity of the river in all seasons, required to provide baseline information against which to assess the impact of an increasing number of exploration and development

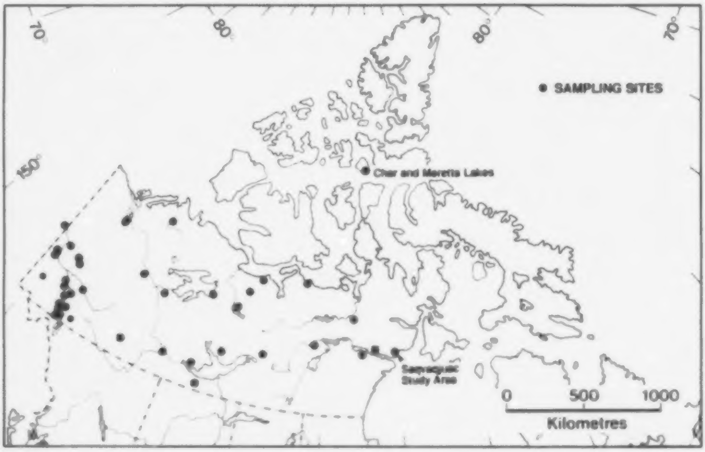

Figure 7.1. Map of Northwest and Yukon Territories showing location of summarized water-quality network stations in N.W.T., distribution of Yukon River stations, "Char" and Meretta Lakes study area at Resolute Bay and the Saqvaqjuac study area.

activities. These concerns resulted in an expansion of the chemical program to include trace metals and nutrients in addition to the physical and major ion parameters of the earlier studies. Data from these stations for selected water-quality parameters are summarized in Table 7.1. The complete data set is published periodically by the WQB (e.g., Canada, Environment Canada, 1976, 1980, 1982, 1984a, 1985 and 1986) while data for the Mackenzie River for the period 1960 to 1979 have been tabulated (Canada, Environment Canada, 1981) and summarized in a report of the Mackenzie River Basin Committee (MRBC, 1985).

It is possible, based on these data, to aggregate the rivers into three loosely-associated groups. The first, the Mackenzie basin rivers, is characterized by higher total dissolved solids (TDS), specific conductance (SC), pH, total alkalinity (Ttl. Alk.) and non-filterable residues (NFR) than the rivers in the other two groups. Included in the Mackenzie basin are the Anderson, Flat, Hay, Liard and Peel Rivers and of course the three Mackenzie River stations. The second group includes the rivers draining the tundra immediately underlain by the Precambrian Shield and the tundra on the glacio-lacustrine clays along the north shore. This group includes the Back, Coppermine, Ellice, Kazan, Quoich, Thelon, and Lockhart Rivers. This group is characterized by the lowest TDS, SC, pH, Ttl. Alk. and NFR of all the rivers sampled in the Northwest Territories. The third group is a transition group and includes the Camsell, Great Bear, Cameron and Ecstall Rivers. The other parameters summarized in Table 7.1 are more variable and cannot be differentiated in this manner. Nevertheless, these data indicate the natural condition of these waters that are characterized by low nutrient

station	period of sampling	total phosphorus (μg·L^{-1}) (15406L)	dissolved phosphorus (μg·L^{-1}) (15103L)	dissolved sulphate (mg·L^{-1}) (16306L)	dissolved magnesium (mg·L^{-1}) (12102L)
Anderson River (1)	69/08/24 to 88/10/26	22±22(23) L3,14,40	5±3(39) L3,3,9	37±23(46) 13,29,69	14±5.4(39) 7.5,12,21
Back River (2)	69/09/12 to 88/10/14	14±15(20) L3,L3,4	3±2(30) L3,L3,4	1.6±1.1(39) 1.0,1.6,2.0	0.8±0.3(32) 0.5,0.8,1.0
Camsell River (3)	69/08/25 to 88/11/14	9±7(23) 3,8,16	4±2(35) L3,L3,6	9.8±2.8(45) 7.0,10,13	5.8±0.9(37) 4.6,5.7,6.8
Coppermine River (4)	72/07/05 to 88/08/08	11±21(30) L3,3,17	4±2(30) L3,3,17	2.1±4.5(50) 1.0,1.4,2.2	0.5±0.1(42) 0.3,0.5,0.6
Ellice River (5)	69/09/06 to 88/09/06	18±14(20) 6,15,30	5±3(28) L3,3,8	2.2±1.2(34) 1.0,2.0,3.3	1.0±0.6(28) 0.6,0.8,1.7
Flat River (6)	72/05/26 to 88/11/30	52±65(28) 7,24,143	7±13(40) L3,3,12	25±10(63) 14,23,39	9.5±3.3(43) 4.8,9.8,13
Great Bear River (7)	69/08/06 to 88/09/22	13±18(35) L3,5,38	4±5(52) L3,L3,6	15±4.4(61) 12,14,16	7.3±1.6(57) 6.1,7.0,8.1
Hay River (8)	68/01/09 to 88/12/13	139±208(67) 22,74,280	28±17(67) 9,28,46	82±45(69) 40,70,128	16±8.6(68) 7.8,14,25
Kazan River (9)	69/09/15 to 88/09/29	28±55(21) 3,8,49	L3(27) L3,L3,3	1.7±2.4(38) 0.9,1.2,2.0	1.0±0.2(31) 0.7,1.0,1.2
Liard River (10)	60/08/10 to 88/12/01	219±416(67) 4,66,583	17±63(66) L3,4,17	30±9.4(75) 18,30,41	11±3.6(73) 6.6,11,16
Mackenzie (11) R. at mouth	60/06/08 to 88/10/13	80±102(56) 7,34,202	6±5(93) L3,4,15	33±10(86) 24,31,39	9.5±3.4(127) 6.9,8.9,13
Mackenzie (12) R. at Norman Wells	60/05/25 to 88/11/02	83±110(45) 3,35,288	6±6(62) L3,3,13	24±4.5(66) 20,23,27	7.9±1.0(91) 6.6,7.9,9.2
Mackenzie (13) R. at Fort Providence	60/07/26 to 88/11/23	39±42(34) 13,24,77	7±3(46) 3,7,12	25±10(56) 19,23,30	7.2±0.9(59) 6.1,7.2,8.2
Peel River (14)	60/06/08 to 88/10/13	146±255(58) 6,36,410	7±10(60) L3,5,12	46±10(60) 33,46,57	13±3.4(81) 7.9,14,17
Quoich River (15)	70/07/07 to 88/10/12	17±23(16) L3,12,25	3±1(26) L3,L3,4	1.7±1.1(35) 1.0,1.5,2.5	0.4±0.2(28) 0.2,0.4,0.6
Thelon River (16)	69/09/13 to 88/10/03	19±25(17) 4,12,25	3±1(25) L3,L3,3	1.9±1.4(32) 1.0,1.3,2.9	1.7±0.5(26) 1.2,1.7,2.2
Tree River (17)	69/07/10 to 88/08/10	14±13(31) L3,9,27	7±14(41) L3,L3,10	1.9±1.2(49) 1.0,1.8,2.9	4.3±0.8(42) 3.4,4.3,5.1
Lockhart River (18)	69/06/12 to 88/09/07	7±4(26) L3,5,13	4±2(36) L3,L3,5	1.9±2.4(43) 1.0,1.4,2.0	1.3±3.8(35) 0.5,0.6,0.8
Cameron River (19)	84/05/02 to 88/09/28	23±52(29) 3,7,14	4±5(29) L3,3,5	3.4±1.0(29) 2.3,3.2,4.5	2.1±0.3(29) 1.8,2.1,2.5
Exstall River (20)	75/06/21 to 85/06/12	52±36(10) 5,46,84	4±2(10) 3,3,4	0.8±0.9(11) L.2,L.2,2.2	3.6±0.7(10) 1.8,3.7,4.0

Table 7.1. Summary of selected water-quality data for N.W.T. stations routinely sampled by Water Quality Branch, Environment Canada. Alphanumeric number in parenthesis in heading identifies parameter code. [mean ± standard deviation (number of samples), 10th, 50th, 90th percentiles].

station	period of sampling	dissolved chloride (mg·L^{-1}) (17206L)	dissolved calcium (mg·L^{-1}) (20103L)	iron extractable (mg·L^{-1}) (26304P)	total zinc (µg·L^{-1}) (30005P)
Anderson River (1)	69/08/24 to 88/10/26	27±24(40) 5.0,16,60	34±11(40) 21,31,50	.28±.46(41) .04,.10,.58	2.7±3.6(23) 0.9,1.3,4.6
Back River (2)	69/09/12 to 88/10/14	1.1±0.6(34) 0.6,0.9,1.7	1.6±0.6(32) 0.9,1.5,2.0	.09±.08(39) .03,.06,.21	12±17(19) 1.0,3.0,25
Camsell River (3)	69/08/25 to 88/11/14	2.2±0.7(36) 1.6,2.1,2.6	14±1.8(36) 11,14,16	.05±.03(47) L.04,L.04,.08	1.7±1.5(16) 0.5,1.0,4.3
Coppermine River (4)	72/07/05 to 88/08/08	0.3±0.1(43) 0.2,0.3,0.4	1.1±0.6(42) 0.5,1.1,1.4	.05±.03(47) L.04,L.04,.09	3.5±3.6(21) 1.0,2.0,5.9
Ellice River (5)	69/09/06 to 88/09/06	3.0±1.8(29) 1.5,2.5,4.7	1.5±0.9(28) 0.5,1.3,1.9	.31±.39(34) .08,.21,.53	4.9±2.8(15) 2.0,4.0,8.4
Flat River (6)	72/05/26 to 88/11/30	0.8±1.5(43) 0.2,0.4,1.2	47±13(43) 29,51,61	1.4±1.9(60) .07,.51,3.6	36±35(51) 7.1,25,79
Great Bear River (7)	69/08/06 to 88/09/22	4.8±1.0(53) 4.0,4.9,5.6	17±3.1(57) 13,17,19	.13±.30(62) L.04,.04,.22	2.0±2.0(29) 0.5,1.0,5.0
Hay River (8)	68/01/09 to 88/12/13	5.9±5.5(71) 2.2,4.1,11	54±27(68) 31,47,85	2.4±2.6(80) 0.6,1.7,5.4	12±11(38) 3.3,7.0,31
Kazan River (9)	69/09/15 to 88/09/29	0.7±0.5(33) 0.5,0.6,0.9	2.7±0.6(31) 1.9,2.6,3.3	.05±.04(36) L.04,L.04,.08	5.1±7.6(18) 0.5,3.0,7.8
Liard River (10)	60/08/10 to 88/12/01	1.4±1.0(74) 0.5,1.0,3.0	41±11(67) 28,40,57	2.7±4.6(95) .09,.70,7.5	16±21(40) 2.0,7.8,39
Mackenzie (11) R. at mouth	60/06/08 to 88/10/13	9.4±4.5(84) 4.5,8.2,14	35±5.2(88) 29,34,41	2.0±2.8(105) .12,.93,5.3	11±13(47) 1.0,5.4,23
Mackenzie (12) R. at Norman Wells	60/05/25 to 88/11/02	8.5±2.2(69) 5.5,9.1,11	27±3.7(60) 23,27,31	1.3±2.9(63) .04,.27,2.9	13±26(37) 1.6,4.0,28
Mackenzie (13) R. at Fort Providence	60/07/26 to 88/11/23	4.4±2.2(50) 1.9,4.1,7.0	33±4.8(59) 27,33,39	.38±.39(81) .08,.26,.70	8.8±8.1(23) 2.5,5.5,20
Peel River (14)	60/06/08 to 88/10/13	2.6±1.6(60) 1.2,2.1,4.2	43±9.2(74) 28,44,53	2.7±4.9(55) .07,.65,8.1	23±32(38) 1.6,7.6,73
Quoich River (15)	70/07/07 to 88/10/12	1.2±1.0(30) .5,1.1,2.0	1.0±0.5(28) 0.2,1.2,1.8	.11±.15(35) .03,.07,.17	15±19(16) 2.4,5.5,35
Thelon River (16)	69/09/13 to 88/10/03	4.3±2.4(29) 1.8,3.2,7.5	3.6±1.1(26) 2.0,3.8,4.8	.09±.05(33) .04,.07,.15	8.3±9.1(17) 1.4,5.4,19
Tree River (17)	69/07/10 to 88/08/10	1.7±0.9(43) 1.0,1.4,2.8	8.2±1.4(42) 6.8,7.9,10	.93±2.01(48) .10,.25,1.8	14±8.7(22) 2.0,9.3,30
Lockhart River (18)	69/06/12 to 88/09/07	2.6±12(37) 0.3,0.4,0.6	3.2±9.8(35) 1.0,1.4,1.9	.03±.01(38) L.04,L.04,.05	1.8±1.4(23) 0.7,1.3,3.0
Cameron River (19)	84/05/02 to 88/09/28	2.0±0.5(29) 1.6,2.0,2.6	7.2±1.1(29) 6.2,6.8,8.2	.04±.02(27) L.04,L.04,.05	1.3±1.0(29) 0.7,1.0,2.0
Exstall River (20)	75/06/21 to 85/06/12	0.9±0.4(10) 0.4,0.8,1.4	8.0±1.3(10) 4.9,8.2,9.0	.90±1.2(14) .04,.67,1.6	8.3±8.5(10) 3.0,5.0,13

Table 7.1. (cont'd) Summary of selected water-quality data for N.W.T. stations routinely sampled by Water Quality Branch, Environment Canada. Alphanumeric number in parenthesis in heading identifies parameter code. [mean ± standard deviation (number of samples), 10th, 50th, 90th percentiles].

station	period of sampling	nonfilterable residue (mg·L^{-1}) (10401L)	ttl. organic carbon (mg·L^{-1}) (06002L)	ttl. N (µg·L^{-1}) (07603L)	ttl. dissolved N (µg·L^{-1}) (07651L)
Anderson River (1)	69/08/24 to 88/10/26	28±74(46) L1,3,70	8.2±3.4(35) 5.0,7.3,12	400±120(39) 270,390,490	350±80(40) 220,350,440
Back River (2)	69/09/12 to 88/10/14	4±5(30) L1,2,8	3.3±1.2(25) 2.2,2.9,4.8	350±190(29) 130,320,580	250±160(31) 100,180,470
Camsell River (3)	69/08/25 to 88/11/14	L1 (37) L1,1,2	4.2±0.6(28) 3.4,4.2,4.9	210±60(32) 150,190,240	180±60(36) 130,170,220
Coppermine River (4)	72/07/05 to 88/08/08	L1 (38) L1,L1,2	2.3±0.5(32) 1.9,2.3,2.5	170±50(37) 140,160,200	150±40(40) 130,140,170
Ellice River (5)	69/09/06 to 88/09/06	10±11(33) 1,6,15	3.7±1.0(20) 2.4,3.7,5.0	240±90(24) 150,220,320	170±90(28) 100,140,240
Flat River (6)	72/05/26 to 88/11/30	104±213(56) L1,27,264	3.9±5.1(29) 1.1,2.1,7.2	280±240(37) 110,200,550	170±90(40) 80,150,210
Great Bear River (7)	69/08/06 to 88/09/22	8±25(60) L1,L1,14	2.2±0.9(42) 1.5,2.0,3.5	270±80(48) 210,250,300	240±60(52) 200,230,280
Hay River (8)	68/01/09 to 88/12/13	92±188(83) 2,16,257	29±8.4(54) 18,27,41	1130±460(61) 700,1040,1550	830±260(66) 530,810,1200
Kazan River (9)	69/09/15 to 88/09/29	2.0±1.0(29) L1,1,3	4.1±1.0(24) 2.9,3.9,5.3	320±180(26) 150,250,640	260±150(28) 130,210,500
Liard River (10)	60/08/10 to 88/12/01	212±391(79) 1,47,677	7.9±8.6(50) 2.1,4.2,16	480±740(59) 170,240,920	170±60(63) 100,170,230
Mackenzie (11) R. at mouth	60/06/08 to 88/10/13	132±211 (110) 3,44,396	7.8±5.1(60) 3.7,5.9,12.5	400±250(70) 220,300,690	210±80(91) 150,200,270
Mackenzie (12) R. at Norman Wells	60/05/25 to 88/11/02	130±323 (113) 1,22,329	6.9±7.2(49) 2.7,4.3,12	420±420(58) 220,290,690	230±70(62) 160,230,300
Mackenzie (13) R. at Fort Providence	60/07/26 to 88/11/23	16±28(89) 3,9,28	11±3.8(35) 6.3,11,16	430±160(40) 270,410,600	340±120(43) 200,320,480
Peel River (14)	60/06/08 to 88/10/13	158±317(75) 1,39,491	7.6±9.7(51) 1.1,3.8,20	510±640(58) 190,240,1110	200±90(61) 130,190,240
Quoich River (15)	70/07/07 to 88/10/12	4.0±9.0(25) L1,1.0,7.0	2.5±1.0(22) 1.4,2.3,3.8	270±190(25) 110,200,480	200±150(26) 70,150,390
Thelon River (16)	69/09/13 to 88/10/03	3.0±2.0(25) 1.0,2.0,5.0	3.5±0.7(46) 27,41,64	330±190(23) 150,240,620	270±170(26) 110,190,510
Tree River (17)	69/07/10 to 88/08/10	20±42(51) 1.0,6.0,37	2.1±0.8(32) 1.3,1.8,2.9	200±120(37) 80,160,370	140±80(42) 70,120,250
Lockhart River (18)	69/06/12 to 88/09/07	1.0±1.0(34) L1,L1,2	2.1±0.5(27) 1.5,2.1,2.7	130±30(30) 90,130,170	120±20(35) 90,120,140
Cameron River (19)	84/05/02 to 88/09/28	1.0±1.0(29) L1,1.0,1.0	6.8±0.6(26) 6.3,6.7,7.6	300±40(29) 240,290,340	260±50(29) 190,260,320
Exstall River (20)	75/06/21 to 85/06/12	62±51(10) 2.0,42,118	4.5±1.1(10) 2.3,4.6,5.7	270±80(10) 200,250,310	180±30(10) 150,170,210

Table 7.1. (cont'd) Summary of selected water-quality data for N.W.T. stations routinely sampled by Water Quality Branch, Environment Canada. Alphanumeric number in parenthesis in heading identifies parameter code. [mean ± standard deviation (number of samples), 10th, 50th, 90th percentiles].

station	period of sampling	ttl. diss. solids ($mg \cdot L^{-1}$) (00201L)	specific conductance ($\mu S \cdot cm^{-1}$) (02041L)	pH (lab) (pH units) (10301L)	ttl. alkalinity ($mg \cdot L^{-1}$) (10101L)
Anderson River (1)	69/08/24 to 88/10/26	213±104(34) 103,167,360	356±169(56) 182,291,597	7.8±1.1(55) 7.6,7.9,8.1	102±34(29) 65,93,145
Back River (2)	69/09/12 to 88/10/14	10±3(25) 7,10,15	21±7(53) 14,19,32	6.7±1.3(52) 6.5,6.7,7.1	5.0±1.4(32) 3.3,4.9,6.7
Camsell River (3)	69/08/25 to 88/11/14	64±11(29) 49,65,76	122±22(62) 96,120,144	7.6±1.2 7.4,7.7,8.0	45.4±6(42) 38,45,54
Coppermine River (4)	72/07/05 to 88/08/08	7±1(34) 5,7,8	14±3(61) 12,13,15	6.8±1.4(62) 6.5,6.8,7.4	3.8±6.1(35) 1.6,3.0,3.8
Ellice River (5)	69/09/06 to 88/09/06	14±7(19) 9,11,26	29±14(44) 16,26,49	6.7±1.8(42) 6.4,6.8,7.6	4.5±2.5(23) 2.1,3.0,7.5
Flat River (6)	72/05/26 to 88/11/30	169±58(32) 104,185,229	312±91(52) 198,341,400	7.9±1.0(51) 7.6,7.9,8.2	127±36(18) 77,122,170
Great Bear River (7)	69/08/06 to 88/09/22	85±8(43) 74,86,93	162±20(74) 132,165,180	7.8±0.9(74) 7.6,7.9,8.1	55±7.2(39) 47,56,61
Hay River (8)	68/01/09 to 88/12/13	264±140(55) 142,222,408	449±202(88) 254,381,742	7.8±1.3(92) 7.5,7.9,8.3	144±84(51) 59,116,253
Kazan River (9)	69/09/15 to 88/09/29	14±4(24) 9,13,18	26±7(52) 20,24,35	6.8±1.1(52) 6.6,6.9,7.3	8.8±2.4(31) 6.1,8.0,13
Liard River (10)	60/08/10 to 88/12/01	169±47(51) 104,159,230	296±85(102) 191,282,411	7.9±0.8(106) 7.7,8.0,8.2	119±39(66) 74,111,171
Mackenzie (11) R. at mouth	60/06/08 to 88/10/13	153±21(61) 124,152,181	275±46(190) 221,272,332	7.7±2.(193) 7.6,7.9,8.1	92±12(130) 75,92,105
Mackenzie (12) R. at Norman Wells	60/05/25 to 88/11/02	123±25(49) 106,125,139	237±50(189) 201,231,266	7.9±1.(188) 7.7,7.9,8.2	78±8.1(150) 68,77,86
Mackenzie (13) R. at Fort Providence	60/07/26 to 88/11/23	138±19(33) 119,137,156	252±65(98) 211,242,294	7.8±2.(102) 7.7,8.0,8.3	88±17(75) 71,86,110
Peel River (14)	60/06/08 to 88/10/13	183±46(54) 111,200,225	328±79(79) 231,337,406	3.5±7.1(80) 7.6,7.9,8.2	118±25(40) 86,118,150
Quoich River (15)	70/07/07 to 88/10/12	8.0±3.0(22) 4,9,11	21±30(46) 8,13,30	5.2±3.9(45) 6.2,6.5,7.0	2.6±1.5(26) L1,2.0,4.6
Thelon River (16)	69/09/13 to 88/10/03	21±7.0(20) 13,21,30	44±16(46) 27,41,64	6.8±1.3(46) 6.5,6.9,7.2	11±3.0(27) 7.5,9.7,16
Tree River (17)	69/07/10 to 88/08/10	41±6.0(31) 33,41,49	79±14(65) 66,75,97	7.6±1.3(65) 7.3,7.7,7.9	35±6.5(39) 29,33,43
Lockhart River (18)	69/06/12 to 88/09/07	9.0±3.0(25) 7.0,8.0,13	17±3.0(54) 14,16,20	6.9±1.5(55) 6.7,7.0,7.4	4.0±1.1(28) 2.1,4.0,4.9
Cameron River (19)	84/05/02 to 88/09/28	32±5.0(26) 28,32,38	68±8.0(29) 58,66,76	7.4±1.1(29) 7.1,7.5,7.7	25±2.6(25) 22,24,28
Exstall River (20)	75/06/21 to 85/06/12	37±5.0(10) 26,38,40	53±27(14) 10,69,75	7.1±2.2(14) 6.5,7.6,7.7	3.1(4) __3.0,__

Table 7.1. (cont'd) Summary of selected water-quality data for N.W.T. stations routinely sampled by Water Quality Branch, Environment Canada. Alphanumeric number in parenthesis in heading identifies parameter code. [mean ± standard deviation (number of samples), 10th, 50th, 90th percentiles].

concentrations, generally low organic carbon, and low concentrations of dissolved ions as well as little iron and zinc.

Several points should be noted regarding these data. First, logistical difficulties mean that only limited samples are collected annually. For example, the Liard River which has been sampled for nearly 30 years has only something of the order of 60 to 70 samples, reflecting a sampling plan intended to provide 3 to 4 samples per year. Thus, the samples may not necessarily be representative of all flow regimes even after many years of sampling. This is especially true for spring flow when, for various reasons, it can be extremely difficult to collect samples. Second, the parameter list does not include many of the environmental contaminants that are currently of concern throughout the rest of Canada such as hydrocarbons, organochlorine pesticides and PCBs. This is in part due to a belief held until recently that minimal use of pesticides and industrial compounds in the North meant that these contaminants would not be present in detectable quantities. The emphasis in this monitoring program is on providing "baseline" data for impact assessment of anthropogenic activities such as mineral extraction (Section 1.5.2.3.). It is appropriate to note that despite the "baseline" objective, this network does not include the High Arctic where extensive exploration and mining activities exist. Not only is there no ongoing water-quality network there, but there has apparently been no further synthesis of water chemistry data for the islands published since that of Thomas (1964). This is not attempted here as it would require a review of published and unpublished documents to assemble data from the disparate research and monitoring activities.

7.2.3. Mackenzie River Basin

The Mackenzie River system is the largest in Canada, flowing nearly 4000 km from the Rocky Mountains in British Columbia to the Beaufort Sea. Fully 52% of the 1.75 x 10^6 km^2 drainage basin is located north of the 60th parallel. Since early settlement over 5000 years ago, the river has played a dominant role in the activity of the region, continuing through the fur trade period to the present when it became a corridor for transportation of goods including products from the expanding oil fields both on the Mackenzie River and in the Beaufort Sea (see Section 1.5.1.). Thus, the attention directed towards the water chemistry of the river is understandable.

Early reports that included information on the chemistry of the Mackenzie River basin include Miller (1947), Rawson (1947), Thomas (1957, 1964) and Livingstone (1966) (see Section 7.2.1.). However, since then a number of studies have focused on this river. Although a few stations in the Mackenzie River Basin have been sampled regularly since the early 1960s (see Table 7.1), most of the sampling has taken place since 1969. As noted above, network data for the period 1960-1979 were reported by the WQB (Canada, Environment Canada, 1981) and summarized for the MRBC (1985).

Nearly 20 years had passed since Miller's (1947) study of the fisheries of Great Bear Lake before Johnson (1975a, 1975b) commenced his study. His objectives were to describe the physical and biochemical limnology of the eighth largest (31,153 km^2), and by far the most northerly, of the world's major lakes, between 1963 and 1965. Although the lake morphometry is complex, Johnson (1975a) states that water from all areas of the freely-circulating basin shows remarkable chemical uniformity which he says is due to thorough mixing during the estimated 124-year residence time, the low degree of biological activity, and the absence of strongly-developed stratification. TDS varied between 78 and 81 ppm, pH was virtually constant at 7.8 - 7.9, and the dominant constituent was bicarbonate at between 64 and 68 ppm. The oxygen in the lake water was always near saturation, even in late winter, at all depths. At the end of winter, the lake shows strong inverse stratification with the cold water extending to 200-m depth. Below 200 m the water is isothermal at 3.52°C and has a uniform level of oxygen saturation suggesting circulation at these depths throughout the winter. The absence of an anoxic layer at the water/sediment interface reflects the extremely low productivity (ultra-oligotrophic) which is confirmed by the high degree of transparency (20 - 30 m secchi-disc depth in open waters). Johnson (1975a) attributes the low productivity more to the low nutrient status of the lake, which has a relatively small drainage basin compared to lake surface area (5:1), than to the low water temperatures which in August are between 4° and 5°C.

Evidently, the only published research on Great Slave Lake since that of Rawson (1950) is a survey of heavy metals in bottom sediments by Allan (1979). The surface sediments were found not to be enriched in Pb and Zn. The central west basin may show As contamination from gold mining activities on the north shore. Both As and Zn were found to decline from relatively high concentrations near tailing-pond effluents of gold mines in the Yellowknife Bay area indicating a possible source for the offshore contamination (Mudroch *et al.*, 1989).

The growing interest in hydrocarbons in the basin during the late 1960s resulted in increasing scientific activity. From late-July to mid-August 1969, a suite of 101 surface samples was taken from throughout the basin and analysed for 22 major and minor inorganic chemical components. These data were supplemented by data from the Canadian Shield part of the basin (Reeder *et al.*, 1972). Factor analyses were used to investigate relationships between the inorganic chemistry of these waters and the physical environment of the river. The results confirmed that the main control on the inorganic composition of the surface waters was the bedrock characteristics of the sub-basins. Aliquots of the samples collected by Reeder *et al.* (1972) were provided to Peake *et al.* (1972) for analyses of amino acids and hydrocarbons. In addition, six sediment samples from mud flats were analysed for saturated hydrocarbons. The amount of total amino acids appeared to be nearly constant throughout the basin while the composition seemed to be simply a reflection of the composition of amino acids in the soils of the drainage basin. The portion associated with suspended particulate matter was about 76% while the dissolved amino acids were frequently in a combined form (>50%). Hydrocarbons were present in trace quantities in only 17 water samples while sediment samples were found to contain from 0.5 to 9.4 $\mu g \cdot g^{-1}$ of dry sediment of normal alkanes (C_{14} to C_{34}). Unfortunately, the sediment samples were not texturally uniform and this accounted for differences among samples as much as did sample location.

Total annual loads of acids and normal alkanes from the Mackenzie River system to the Beaufort Sea were estimated to be 5.6 - 6.1 x 10^{10} g and 2.4 x 10^8 g, respectively.

The Canadian Government's Task Force on Northern Oil Development during the 1970s encouraged additional work in the Mackenzie River basin. Reeder (1973) and Reid *et al.* (1974) reviewed the water-quality data for the basin network that had been established and expressed concern that the existing high levels of nutrients and heavy metals in many parts of the basin might be further increased by bank and bed material erosion due to pipeline and road construction. Reid *et al.* (1975) sampled bank material from seven rivers, identified by Reid *et al.* (1974) as potentially sensitive, during periods of surface run-off. Carbon, phosphorus, nitrogen, iron, copper, zinc and specific conductance were considered as the most likely to be affected. It was concluded by Reid *et al.* (1975) that construction would affect local water quality, but would generally not impair the quality of water in the Mackenzie River basin.

Brunskill *et al.* (1973a, 1973b) considered the effects of construction on the river ecology, as opposed to just water chemistry. Their research concluded that lakes and streams with small watersheds were likely to be affected by terrain disturbances to a greater extent. Further, the disturbances on small systems were more likely to be longer-term. Rivers and streams with lower water velocities could not transport the increased supply of sediment from disturbances. This resulted in deposition of fine silts and clays on the gravel substrates which were apparently deleterious to the growth and reproduction of zoo- and phyto-benthos.

Additional investigation of Mackenzie River basin sediments showed that some of the larger rivers had high transport rates of total phosphorus (TP) and nitrogen (TN), but that a large proportion of both was in the particulate form and thus not likely readily available for biological utilization (Brunskill *et al.*, 1975). Wagemann *et al.* (1977) determined the trace metal composition and reactivity to Beaufort Sea water of sediment samples from the basin. Fe, Mn, Co and Cr were found to be in forms separable from the particulate phase by mild leaching. When reacted with Beaufort Sea water, both Fe and Mn were released from the sediments to a small degree while Cu, Pb and Zn were adsorbed by sediments.

Concern with further development within the basin prompted research on the ecology of freshwater habitats in the Mackenzie Delta (Fee *et al.*, 1988). Samples were collected during ice-free periods of 1985 and 1986 on a number of lakes. The ice-free season phytoplankton-production rates were low but comparable to other arctic lakes (Kalff and Welch, 1974). Low transparency, limited nutrients, and shallow depths all contribute to the low aerial production rates.

In summary, there is an extensive water-chemistry data base available for the large and complex Mackenzie River basin, but there still seems to be only a limited understanding of its aquatic ecosystem. Some of the requirements to overcome these deficiencies will be discussed below.

7.2.4. Yukon River Basin

The Yukon River basin, with a total Canadian drainage area of 294×10^3 km^2, received considerably less attention than the Mackenzie River basin until recently. Historically, this may be due in part to the absence of large natural lakes within the Yukon basin, such as were studied for fisheries exploitation in the Mackenzie basin. Recent Yukon River basin studies have tended to be more systematic than those of the Mackenzie, but their inconsistency means there is no long-term data base.

Early water-quality work is reviewed in Section 7.2.1. and is limited to Wynne-Edwards (1947a) and Thomas (1957). Water-quality monitoring data for the period 1961 to 1973 were reported by the WQB (Canada, Environment Canada, 1975). Seventeen of the stations reported here (see Figure 7.1) have been sampled 3 to 4 times per year at least since 1969 for major ions, nutrients, and selected metals and trace elements. Evidently, this basin-wide monitoring did not continue as the next reported basin study was for a period of 13 months at 21 stations during 1982 and 1983 (Jack *et al.*, 1983). Sampling frequency varied from weekly to seasonally, depending on the station, and analyses included physical parameters, nutrients, ions and trace metals. This study observed that the waters were soft, slightly alkaline with a pH between 7.0 and 8.3 and that only the concentrations of copper and lead exceeded aquatic life objectives[b]. A more intensive investigation of Hg on the Pelly River indicated concentrations near or below detection limits and not in excess of guidelines.

Whitfield (1983) investigated cross-sectional heterogeneities and temporal variations for four sites on the Yukon River for a number of water-quality variables between 1974 and 1979 in order to estimate the optimum frequency and density of sampling required for long-term monitoring. Most sites and variables could be represented by a single weekly sample. These conclusions led to a study in 1982 and 1983 with weekly sampling at seven sites and at two-week intervals at three sites (Whitfield and Whitley, 1986). The classification system which considered water-quality variable responses to discharge was consistent with physical mechanisms postulated for each group and may be useful in characterizing other northern waters with more limited water-chemistry information.

The long periods of ice cover on northern rivers could result in severe dissolved oxygen (DO) depressions during the winter season. This concern led to investigations in the Yukon River basin (Schreier *et al.*, 1980). Samples were collected six times from the fall of 1977 through to the spring of 1978 on Swift River in the southern Yukon and on Ogilvie River in the northern Yukon. DO and SC were determined in the field and analyses for physical parameters, nutrient, metals and ions were conducted in the WQB laboratory. This study confirmed severe DO depressions throughout the basins with lowest levels in late winter. Open water did not necessarily aerate the river as it was frequently linked to ground-water inflows which often exhibited low oxygen concentrations. Recovery from low DO was rapid with ice break-up.

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[b] *"Aquatic life objectives" for copper and lead are 0.002 and 0.005 mg·L^{-1} respectively. Water-quality use objectives are designated concentrations of constituents that when not exceeded will protect a prescribed water use with an adequate degree of safety (Canada, Environment Canada, 1979). In this case, the protected use is aquatic life.*

The Takhini River, a ground-water dominated system, and the Nordenskiold River, rich in organics and with many bogs and marshes, were sampled to further investigate DO under ice (Whitfield and McNaughton, 1986). DO decreased very rapidly in November and continued to decline in both rivers until February or March. The Nordenskiold River became more depressed than the Takhini River, but both rivers showed DO recoveries during April under full ice cover. The fact that DO increased while ice was present, concurred with the conclusion of Albright *et al.* (1980b) that the Ogilvie and Swift Rivers had greatest microalgal and bacterial biomasses in spring and summer, likely controlled by light and dissolved organic carbon. DO increases under ice had not previously been reported, probably due to the absence of detailed measurements just prior to ice-out. Prowse and Stephenson (1986) confirmed the controlling influence of ice and snow cover on lake light penetration and thus DO in a temperate Ontario lake. Snow cover attenuates light much more than does either black or white ice.

In contrast to the Northwest Territories, several studies have investigated the limnology of Yukon lakes extensively. The results for 91 lakes, studied from 1960 to 1978, were reported by Lindsey *et al.* (1981). This report presented the physical and biological limnology of these lakes, including their chemical composition (major ions), chlorophyll, oxygen, transparency and temperature regimes. Four of these lakes (Atlin, Laberge, Marsh, and Tagish) were investigated in greater detail in 1982 and 1983 to better understand the processes controlling their limnology in order to assess the effect of potential developments within the basin (Kirkland and Gray, 1986). These authors concluded among other things that the biological productivity in the lakes is nutrient-limited and that total phosphorus is not a good indicator of biologically-available phosphorus due to large quantities of suspended sediments. The optimal time for algal uptake of nitrate is suggested to be between snowmelt and ice-out as there is adequate light at this time and an absence of wind-induced mixing. These four large lakes showed TP concentrations about one half, and TN concentrations less than 20% of those reported by Shortreed and Stockner (1986) for 19 smaller Yukon lakes studied in 1982. The very low concentrations make the use of nutrient ratios for determining trophic state suspect. The study indicated that most of the lakes could be classified as oligotrophic, although several are mesotrophic. Most of these smaller lakes showed slight to severe phosphorus limitation while the absence of blue-green algae in the four lakes studied by Kirkland and Gray (1986) was suggested as indirect evidence that phosphorus remains the limiting nutrient.

Shortreed and Stockner (1986) observed that chlorophyll levels under ice were higher than average summer chlorophyll in five of the ten lakes sampled. Production under ice has also been indirectly measured by DO concentrations in both arctic rivers and lakes (Schreier *et al.*, 1980; Whitfield and McNaughton, 1986; Albright *et al.*, 1980b; and Prowse and Stephenson, 1986). This indication that biological productivity in natural arctic lakes is controlled by light, with maximum productivity commonly occurring in the spring after snowmelt, but still under ice cover, suggests that phosphorus-chlorophyll relationships for predicting trophic state, such as those proposed by Vollenweider (1976) and Ryding and Rast (1989) may have limited application in the North. For example, these relationships frequently use phosphorus concentrations at spring overturn to predict peak summer chlorophyll. This relationship lacked a significant correlation for arctic lakes (Shortreed and Stockner, 1986). Spring peak chlorophyll rather than summer peak chlorophyll may be the important variable in future research into the control of eutrophication in arctic systems.

As in the Mackenzie basin, concern arose in the Yukon basin over the effect of stream-bank erosion on the aquatic quality of the system, due to highway and pipeline construction. Another study on the Ogilvie and Swift Rivers (Albright *et al.*, 1980a) indicated that excavation activities would have minimal effects during and immediately following the freshet since natural sediment concentrations were high, DO levels were high and microbiological activities and biomass would recover quickly. These sub-arctic rivers would be most sensitive to increased erosion during the winter as the already low DO levels would be decreased further and microbial recolonization would be slow.

Although water-quality data in the Yukon do not have as long or consistent a record as in the Mackenzie River system, the more recent studies have focused on a better understanding of natural phenomena important to discerning the potential impacts of anthropogenic activities. This research provides important background for designing future research programs and monitoring strategies throughout the North.

7.2.5. Limnological Research Activities near Resolute, Cornwallis Island and at Saqvaqjuac

As recently as 1970, Kalff stated that "the greatest need in Arctic limnology is for comprehensive studies of lake ecosystems with individuals pooling their efforts" (Kalff, 1970). The absence of data from the Canadian North is demonstrated by the fact that the only arctic lakes for which Kalff could obtain published information regarding primary productivity were from Alaska and Lapland. Fortunately, some excellent progress has been made since that time with comprehensive investigations of two lake systems - "Char" and Meretta Lakes near Resolute on Cornwallis Island and a number of lakes in the Keewatin District near Saqvaqjuac. The location of these two areas is shown in Figure 7.1.

The Char Lake work began about 1969 and continued through 1973 as part of the International Biological Program study to describe, for the first time, the complete annual cycle of physical and chemical events in a high arctic lake (Schindler *et al.*, 1974b). In most years Char Lake is a typical polar lake never rising above 4°C and with a brief summer overturn period of up to 6-weeks duration. The lake is extremely oligotrophic with soluble reactive phosphorus measurements (SRP) consistently less than detection ($<0.7\ \mu g{\cdot}L^{-1}$); nevertheless winter oxygen depletion at the sediment/water interface was considerable, probably reflecting respiration by the benthic community. DO concentrations were greatest just under the ice when it was thickest in late-May (Schindler *et al.*, 1974b). This probably reflects increased respiration in the surface waters in association with increased light penetration (24 hours of sunlight and melting snow) at this time of year (Albright *et al.*, 1980b, Prowse and Stephenson, 1986; Whitfield and McNaughton, 1986). Schindler *et al.* (1974b) also noted that mixing continues after the lake is frozen over as evidenced by the general uniformity of SC with depth. Additionally, there was evidence of freeze-out of solute-rich water which accumulated in the deepest part of the lake during the winter.

The eutrophication of high arctic lakes was also of interest and Schindler *et al.* (1974a, 1974b) estimated the annual phosphorus (P) and nitrogen (N) budgets for Char Lake to be 0.016 and 0.314 $g \cdot m^{-2}$ respectively. These figures were the lowest reported for any lake at that time. The N:P ratios entering the lake were 20:1 suggesting that this lake would respond to additions of phosphorus rather than nitrogen. A more detailed nutrient budget was provided by de March (1975). Run-off, in particular snowmelt, was the main source of nutrients to the lake. Only a very small portion (8%) of the dissolved P was retained in the lake while about 75% of the particulate P and 50% of the dissolved nitrogen were retained. De March (1975) concluded that this high nutrient loss from the lake could be explained by the presence of a dilute meltwater layer which occurred under the ice and which flowed out of the lake before mixing and before nutrients could be utilized by the biota.

Meretta Lake, which is about 3 km from Char Lake, has received untreated effluent from the Resolute base camp (permanent population of about 150) since 1949 and provides an excellent opportunity for a comparative study. Meretta Lake was sampled in a manner comparable to Char Lake between 1969 and 1972 (Schindler *et al.*, 1974a, 1974b). Total annual inputs of P and N to Meretta Lake were determined to be about 0.2 $g \cdot m^{-2}$ and 0.6 $g \cdot m^{-2}$ respectively, or about 10 times more P and 2 times more N than estimated for Char Lake. According to standing crop, these authors classified the lake as mesotrophic whereas such nutrient inputs in temperate lakes would normally create eutrophic conditions. This difference was thought to be due more to an inappropriate N:P nutrient ratio for optimal algal growth rather than to a latitudinal response. Although Schindler *et al.* (1974a) concluded that neither phosphorus nor nitrogen recycled from the sediments, implying that the lake would return quickly to natural conditions once the sewage input was eliminated, they did note that some species native to these waters were declining in numbers or had disappeared due to changing oxygen, food or sediment conditions.

Limnological research into the effect of anthropogenic activities on water quality in the Canadian High Arctic has not continued near Resolute Bay. Rather, a multi-disciplinary research project was started at Saqvaqjuac, near Chesterfield Inlet on the west coast of Hudson Bay (see Figure 7.1), in 1976 as a parallel project within the Arctic to the Experimental Lakes Area in southern Manitoba (Welch, 1985). This is an area of numerous glacial ice-scoured lakes most of which are small (<50 ha). The lakes are cold, monomictic, circulating once annually during the summer with temporary thermal stratification in midsummer. The water is generally soft with total dissolved solids concentrations between 15 and 30 $mg \cdot L^{-1}$, oligotrophic and clear. Water circulation and winter respiration of these lakes has been discussed by Welch and Bergmann (1985a, 1985b) while the chemical limnology has been reviewed in Welch and Legault (1986). This was the first detailed investigation of element input/output in continuous permafrost watersheds. The mean pH of fresh snow was 4.65 and that of summer rainfall was 4.80. The year-to-year variability in solute output from permafrost watersheds was high and determined by the timing and magnitude of the snowmelt run-off. In summary, Welch and Legault (1986) stated that the solute output was low in years with little mid- to late-summer rainfall resulting in the run-off being made up of snowmelt early in the summer before the active layer was well developed. The importance of mid- to late-summer rainfall was demonstrated by its coincidence with high solute outputs. This was also an important factor with respect to lake element loading and retention. If most run-off occurred prior to ice-off, the inflowing water and solutes outflowed from the surface of the lake prior to mixing with the rest of the lake water where biological removal could take place. These lakes tended to have lower DO concentrations than high arctic lakes (Welch, 1974) due to higher respiration rates.

7.3. CURRENT RESEARCH - LONG-RANGE ATMOSPHERIC TRANSPORT AND DEPOSITION

The relatively recent realization that the Canadian arctic aquatic environment is not sufficiently remote to be protected from the effects of long-range atmospheric transport of pollutants from highly populated/industrialized regions has led to a new thrust in arctic research. Because these contaminants are atmospherically transported they are seemingly ubiquitous. The fact that the source regions are international places the protection of the arctic environment in a global context. Thus current and future research in the Arctic must go beyond domestic problems, such as mining and industrial activities, and become interdisciplinary in scope and hemispheric in scale.

Recently, a macro-scale study was undertaken for the purpose of characterizing the aquatic sensitivity of the N.W.T. surface waters to acid deposition through a synthesis of all acceptable surface-water quality data (Anonymous, 1987; Tibbatts *et al.*, 1987). This analysis showed large areas of the eastern mainland underlain by the Canadian Shield as highly sensitive. The Arctic Archipelago was not included in this assessment due to the absence of surface-water chemistry data.

Very recently, contamination by trace organic compounds has come to the forefront of environmental-quality research in the Arctic, as in other parts of the world. In order to place this research in context, it is necessary to very briefly summarize findings from outside the specific area of freshwater quality.

A variety of trace organic contaminants, including PCBs, toxaphene and other organochlorine pesticides have been observed in arctic marine mammals, fish and fauna from as early as 1972 (Holden, 1972) through to the present (Norstrom *et al.*, 1988; Muir *et al.*, 1988; Bidleman *et al.*, 1989).

The dependency of the Inuit population upon natural or "country" foods in their traditional diets has been outlined by Wong (1985) while the presence of trace organic contaminants in the country foods and the appearance of PCBs in blood and breast milk in Baffin Island and northern Quebec communities has been discussed by Kinloch *et al.* (1990) and Dewailly *et al.* (1989). In the Quebec study, the PCB concentrations in milk fat of Inuit women was almost five times higher than Caucasian women and breast-fed babies were estimated to achieve PCB blood level of 150 $\mu g \cdot L^{-1}$ in 3 to 18 months. Dewailly *et al.* (1989) suggest that PCBs could play a role in the impairment of immunity and in the high occurrence of infection among Inuit children.

The understanding of the processes and pathways determining the presence and accumulation of these contaminants in the Arctic is poorly understood. A complete review of the current state of knowledge, albeit seriously limited regarding the sources, occurrence and pathways, is in preparation (Barrie *et al.*, 1990). The relative importance of background marine contamination and atmospheric transport, deposition, snowmelt and run-off to the body burden of marine and freshwater species is unknown. The studies of the arctic haze phenomenon and in particular the work of Carlson (1981), Barrie *et al.* (1981), Rahn and Shaw (1982), and Lowenthal and Rahn (1985) all suggest that the Canadian arctic atmosphere is being subjected to extensive incursions of measurably-contaminated air masses. The history of Canadian observations of arctic aerosol chemistry has been summarized by Whelpdale and Barrie (1982) and Barrie and Hoff (1985). For a short period in the summer of 1984, the Canadian aerosol-chemistry sampling station at Mould Bay was used to measure chlorinated compounds, specifically chlordane and its oxidation/metabolization products. Chlordane was only detected in the gas phase at the $pg \cdot m^{-3}$ level. At about the same time, outside of Canada Oehme and Ottar (1984) were measuring the presence of polychlorinated hydrocarbons in the Norwegian Arctic. Airborne organochlorines were reported for the Canadian High Arctic by Hargrave *et al.* (1988), Patton *et al.* (1989) and Patton and Bidleman (1989) and for toxaphene by Bidleman *et al.* (1989). Until recently there was only limited evidence of the fate of these compounds in the arctic aquatic system.

Several studies have been conducted on trace organic contaminants in arctic snow by Stengle *et al.* (1973), McNeely and Gummer (1984), and Paasivirta *et al.* (1985). These studies reported either no measurable trace organic substances or very low concentrations near the respective analytical detection limits. This research likely led, to some extent, to the dismissal of the atmosphere as an important transport pathway of trace organic contaminants to the Canadian Arctic. The measurement of organochlorine contaminants in the snowpack sampled from a wide geographical area (see Figure 7.2) by Gregor and Gummer (1989) strongly supported the long-range atmospheric transport/deposition hypothesis. They reported that hexachlorocyclohexanes (HCHs) were dominant in the snow, consisting of the pesticide lindane (0.20-4.08 $ng \cdot L^{-1}$) and alpha-HCH (0.43-8.72 $ng \cdot L^{-1}$). Other compounds included alpha-endosulfan, dieldrin, chlordane and heptachlor epoxide at concentrations of $ng \cdot L^{-1}$ and PCBs at between 0.02 and 1.76 $ng \cdot L^{-1}$. The 1986 survey was more-or-less repeated in 1987 and the concentrations and winter deposition rates of selected contaminants are

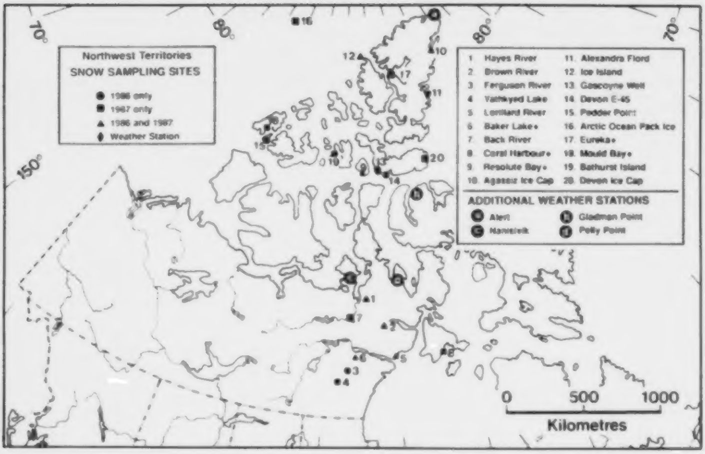
Northwest Territories
SNOW SAMPLING SITES
● 1986 only
■ 1987 only
▲ 1986 and 1987
♦ Weather Station
70°
80°
80°
70°
150°
1 Hayes River
2 Brown River
3 Ferguson River
4 Yathkyed Lake
5 Lorillard River
6 Baker Lake•
7 Back River
8 Coral Harbour•
9 Resolute Bay•
10 Agassiz Ice Cap
11 Alexandra Fiord
12 Ice Island
13 Gascoyne Well
14 Devon E-45
15 Pedder Point
16 Arctic Ocean Pack Ice
17 Eureka•
18 Mould Bay•
19 Bathurst Island
20 Devon Ice Cap
ADDITIONAL WEATHER STATIONS
Alert
Gladman Point
Nanisivik
Pelly Point
0
500
1000
Kilometres

Figure 7.2. Map of Northwest Territories showing the locations of snow sampling sites for 1986 and 1987.

Compound	1986 conc'n. ($pg{\cdot}L^{-1}$)	1986 flux ($ng{\cdot}m^{-2}{\cdot}yr^{-1}$)	1987 conc'n. ($pg{\cdot}L^{-1}$)	1987 flux ($ng{\cdot}m^{-2}{\cdot}yr^{-1}$)
PCB	745 (20-1760)	73	907 (257-1773)	95
alpha-HCH	3185 (430-8120)	285	910 (143-42700)	95
gamma-HCH (lindane)	1160 (220-4446)	104	700 (83-10500)	95
heptachlor epoxide	97 (ND-362)	10	117 (ND-1600)	95
alpha-endosulfan	345 (100-1094)	37	175 (ND-4880)	95
dieldrin	540 (205-1402)	58	400 (ND-4420)	95
HCB	28 (10-70)	3	16 (ND-104)	2
sum-DDT	18 (10-50)	2	93 (ND-1380)	8
sum-CBZ	NA	NA	240 (ND-915)	30

NA = not available ND = Not detectable
HCB = hexachlorobenzene
sum-DDT = sum of all detected DDT compounds and metabolites
sum-CBZ = sum of all detected chlorobenzene compounds targetted

Table 7.2. Median and range of concentrations and calculated winter season deposition to the Arctic snowpack of organochlorines during 1986 (n=12) and 1987 (n=15) surveys *(Barrie et al., 1990).*

reported in Table 7.2. These findings were supported by Erickson *et al.* (1988) and Hargrave *et al.* (1988), working on the Canadian Ice Island located in the Arctic Ocean at about 80°57'N, 97°35'W, in the spring and summer of 1986. These low concentrations are not particularly instructive on their own.

The relative significance of these contaminant concentrations is better illustrated by the calculated atmospheric loadings of selected compounds to the snowpack as shown in Table 7.3. The estimated annual deposition of pesticides to the Canadian Arctic is nearly equal to that of temperate Canada except for British Columbia. The total PCB loadings to the Arctic on the other hand are lower than temperate sites. Although tempting, extrapolation of these loadings to the arctic land mass to estimate total deposition must be avoided until a more extensive data base is available.

compound	**Lake Superior**[b]	**Central Ontario**[a] **(1984)**	**North Saskatchewan** [c]
a-HCH	3.9 - 11	14	2.1 - 6.5
lindane	2.0 -3.7	7.3	.30 - 1.9
dieldrin	.15 -.50	.30	.01 - .12
Σ-DDT[a]	.08	.02	.002 - .01
Ttl.PCBs	1.2 - 4.9	1.6	1.5 - 1.7
HCB	.04 -.05	.08	.003

compound	**South Alberta (1985)**	**South British Columbia**[d]	**Arctic**[e]
α-HCH	3.3	22 - 36	.01 - 2.3
lindane	1.3	4.8 - 6.2	< .01 - 1.0
dieldrin	.002	0.3 - 0.34	0.0 - .25
Σ-DDT[a]	.007	0.3	< .01 - .2[f]
Ttl.PCBs	1.2	1.0	.01 - .3
HCB		.05	< .01

[a] Strachan (1988) reported on p,p'-DDE as opposed to Σ-DDT for the Arctic.

[b] The Lake Superior data are from 1981, 1983 and 1984.

[c] The Cree Lake station, northern Saskatchewan, included data collected in 1984 and 1985.

[d] The Kanaka Creek station, southern B.C., included data from 1985 and 1986.

[e] The arctic data are ranges for the two years, 1986 and 1987, for all sites except Alexandra Fiord.

[f] Represents the range of loadings for 1987 sites only.

Table 7.3. Comparison of annual contaminant deposition in precipitation in non-arctic Canada[a] with Canadian arctic winter (September to May) loadings estimated from concentrations in the snowpack (µg·m^{-2}).

Toxaphene, the most abundant contaminant in marine seals and fish (Muir *et al*., 1988) was also present in the snow from three sites in the Canadian Arctic at between 85 and 1700 pg·L^{-1} (Bidleman *et al*., 1989).

The importance of long-range atmospheric transport and deposition of contaminants to arctic water quality is demonstrated by Nagy *et al*. (1987) who measured hydrocarbons in the water and suspended sediments of the Mackenzie River. These investigators concluded that in view of the distribution of hydrocarbons within the system, the in-stream local sources including natural seeps and escapes from the Norman Wells site would be overwhelmed by diffuse inputs from spring run-off.

Published measurements of organochlorines in Canadian arctic lakes and rivers are, for all intents and purposes, non-existent except for limited data from the Mackenzie River (Erickson and Fowler, 1987). Some data from northern Ontario and Quebec have been published by McCrea and Fischer (1986) and Langlois (1987) respectively. As with snow, hexachlorocyclohexanes are dominant in these waters and it is informative to compare these data with some from Lakes Superior and Ontario and unpublished data for Sophia Lake (75°5'N, 93°40'W) and Lakes Hazen and "Amituk" (81°48'N, 70°55'W) in Table 7.4. The Lake Hazen concentration is the median of three samples collected from 2 m below the bottom of the ice in May, 1989. Concentrations at the 35-m depth (not shown in Table 7.4) are slightly lower. The high concentrations measured at Sophia Lake are greater than the average concentrations of Lakes Superior and Ontario. The elevated concentrations in Sophia Lake are perhaps attributable to the fact that this is a meromictic lake with a very shallow surface layer resulting

isomer	Lake Superior[a]	Lake Ontario[a]	N. Ont. rivers[b]	N. Quebec Hudson B.[c]	N. Quebec Ungava B.[c]	Mackenzie R./Delta[d]	Sophia Lake[e]	Lake Hazen[e]	Lake Amituk[e]
n	19	31	60	87	87	n/a	2	3	1
α-HCH	8.0	4.1	6.4	8.0	5.8	3.0 - 4.9	13.6	0.6	3.3
lindane	1.1	1.3	0.9	1.1	0.8	0.4 - 0.5	1.7	0.2	0.5

[a] from Stevens and Neilson (1989)
[b] from McCrea and Fischer (1986)
[c] from Langlois (1987)
[d] from Erickson and Fowler (1987)
[e] unpublished data from D.J. Gregor based on samples collected 2 m below the ice in May, 1989

Table 7.4. Comparison of hexachlorocyclohexane (lindane and alpha-HCH) concentrations in water (ng·L^{-1}) from the Mackenzie River and Sophia, Hazen and Amituk Lakes, N.W.T. to concentrations in rivers in northern Ontario and Quebec and to Lakes Superior and Huron.

in trapping and minimal dilution of surface run-off. This seems to be supported by the lower lindane and alpha-HCH concentrations at 2-m depth in an unnamed, cold, monomictic lake approximately 7 km from Sophia Lake (here referred to as "Amituk" Lake).

Since a long-term data base does not exist to provide historical records of contaminant deposition in the Arctic, attempts have been made to reconstruct such information from glaciers. The most complete work in this regard is the acidity record of Agassiz Ice Cap on north-central Ellesmere Island where the annual maximum conductivity/acidity of the ice has increased by about 75% between 1956 and 1977 as European SO_2 and NO_x emissions increased (Barrie *et al.*, 1985; Koerner and Fisher, 1982). Stengle *et al.* (1973) conducted some early work on a glacier on Mount Logan, Yukon (5364-m elevation) for the purpose of defining a DDT deposition record. These authors reported that DDT was not present above the detection limit of 5 $ng \cdot L^{-1}$ in any of the 19 snow samples collected at depths of 1 to 15 m below the surface. Based on data published by Gregor and Gummer (1989) this detection limit was too high for the concentrations expected in arctic snow.

The only other work on historical trends of trace organic contaminants is that of Gregor *et al.* (1990). This research, conducted on the Agassiz Ice Cap in north central Ellesmere Island, considered 17 years of annual accumulation including the fresh seasonal snowfall for the winters of 1985/86 and 1986/87. The authors hypothesize that due to re-volatilization during the summer season, the contaminant residue in the glacial snow is only a small fraction (about 10%) of the total contaminant deposition to the glacier during the winter season. Consequently, it is impossible to use glacial residues to determine depositional trends in the Arctic until the physiochemical controls of these compounds in this environment are better understood. This is particularly true for high vapour pressure compounds such as the HCHs. An exception is Total PCB, measured as an Aroclor mixture, of which approximately 75% appears to remain in the glacier. Total PCB residues in the glacier show a decline in annual flux during the 17 years from >300 to about 100 $ng \cdot m^{-2}$. Gregor *et al.* (1990) have attempted to adjust this Total PCB residue trend for inter-annual meteorological variability using ^{210}Pb. These results, shown in Figure 7.3, suggest that the Total PCB flux to the glacier has remained stable since 1976/77 at a level approximately one half of that which prevailed between 1970/71 and 1975/76.

Although limited geographically, temporally and conceptually, the current research activities have demonstrated the widespread presence of toxic chemicals in the North. The contamination of the "country foods" and the threat of ecological damage in the Arctic by chemicals essentially foreign to the region illustrate how far-ranging chemical pollution is and yet we know very little about potential effects on the arctic aquatic environment. Hopefully, the realization that the Arctic is no longer pristine will reinforce the rising sense of responsibility in the developed and industrialized parts of the world to protect the environment from further stress. Unfortunately, due to the delay in recognizing long-range atmospheric transport and deposition of pollutants, the Arctic may be the site of an unplanned and uncontrolled

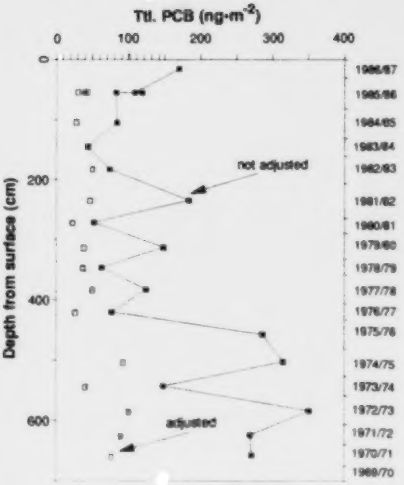

Figure 7.3.
Trend of Total PCB residues in the Agassiz Ice Cap, Ellesmere Is., N.W.T. between 1970/71 and 1986/87 and the adjusted Total PCB residue trend using ^{210}Pb to adjust for inter-annual depositional variability *(Gregor et al., 1990)*.

experiment on the ability of polar ecosystems to withstand, adapt and/or recover from environmental stresses imposed by acid and chemical deposition.

7.4. FUTURE RESEARCH NEEDS

The situation in the Canadian Arctic during the past four decades has changed little since Rawson (1953) stated that "as yet very little has been done in the Arctic". It is noteworthy that he recommended a reconnaissance of lakes in the North over thirty years ago and yet this is still lacking. He also recommended intensive investigations which have only been pursued to a limited extent as described above. In this context, research initiatives must attempt to reproduce for the Arctic the level of understanding presently sought for temperate regions. It is not sufficient to assume that the processes, responses and sensitivity of the northern ecosystems are comparable to temperate ecosystems until the North is better understood. Some of this work was commenced as part of the research undertaken near Resolute and at Saqvaqjuac but it needs to be continued and expanded. The opportunity exists in the North to establish long-term, or even representative, ecosystem research basins which can be protected or managed in a way to permit a comprehensive understanding of local and large-scale environmental impacts and trends through research and monitoring. This opportunity, if it is not to be lost, must be acted upon soon.

It seems trite but necessary to say that there is a need in the Arctic for holistic, systematic, process-response research to be conducted as an umbrella for the often-fragmented activities undertaken to address specific issues or problems. The Saqvaqjuac site, for example, provides an excellent opportunity for commencing intensive, comprehensive investigations of the fate and effect of contaminants in the Arctic. More than seven years of data are available for Saqvaqjuac lakes (Welch and Legault, 1986) and this data base provides an excellent starting point for additional limnological research in the Arctic. As the evidence mounts that long-range atmospheric transport is delivering important quantities of trace contaminants to the Arctic, the understanding provided by these studies will prove to be invaluable groundwork for the study of the fate and effects of, for example, chlorohydrocarbons in the arctic aquatic environment. However, regional complexities have not been addressed and thus there remains a need for comparable work in other geographical regions.

The existing WQB/WRB surface-water monitoring program needs to be expanded in a scientifically-justifiable manner to better represent critical hydrophysical conditions and to better reflect the pollutant spectrum generated from long-range atmospheric transport. Specifically, the monitoring plan warrants evaluation with respect to site selection, sampling frequency and site homogeneity. Some of the information gained from the studies in the Yukon River basin is likely relevant to the Mackenzie River basin and other parts of the Arctic. This needs to be confirmed. Some of the information needs for the Mackenzie River may be forthcoming from ongoing studies of the potential effects of planned upstream forest activities. On the other hand, further consideration should be given to obtaining a longer-term historical water-quality record at a number of sites within the Yukon River basin using existing information with respect to downstream mixing, cross-sectional variability and appropriate sampling frequencies. There is also a clear need to expand the surface-water monitoring program geographically so that the Arctic Archipelago is represented.

As noted above, there has been no systematic sampling program of arctic lakes on a large scale to date and this merits consideration more than ever before to provide "snapshot" data. Aside from the acid-sensitivity mapping there has also been no effort to assemble, evaluate and synthesize data from the numerous sources available and this should be undertaken prior to a macro-scale sampling program.

The need for research on toxic chemicals should be a high priority. Specific activities need to include improved spatial and temporal assessment of deposition, better understanding of the fate of these contaminants in various aquatic systems (e.g., rivers, lakes, glaciers) and a much improved understanding, including both laboratory and field measurements, of the biophysiochemical properties controlling rate functions. Continued investigations of historical deposition trends are critical in order to predict the effects of changing deposition on the environment. Finally, the capability to integrate these data into a model framework for synthesis, remedial-action response prediction and information-gap identification is required. Finally, co-ordinated circumpolar co-operation is essential to complete this work since pollution sources are international.

There will of course be stumbling blocks to achieving these activities in the North. The high logistical costs of research necessitate funding considerably in excess of that normally available for environmental research. There also seems to be a lack of interested and trained scientists to work in the region on a continuing basis. The small community of scientists existing in Canada, at least relative to the area involved, commonly are operating outside of an overall framework for northern research and with no long-term commitment for support.

Additionally, the work is being undertaken largely in isolation nationally and internationally and there is a need for circumpolar co-operation such as proposed under the International Arctic Science Committee (IASC) (Anonymous, 1988). Finally, care has to be taken to avoid a backlash against "southern controlled" research. With the settlement of land claims in the region, there will need to be increased effort in communicating with and working with natives as well as providing a training function to assist the indigenous people in achieving a level of internal capability in scientific research and resource management. This perhaps will be the greatest challenge to undertaking effective water-quality research and monitoring activities in Canada's North in the longer term. Hopefully, the decade of the 1990s will see greater equality and opportunity for northern environmental research in Canada.

7.5. ACKNOWLEDGEMENTS

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8

THE REGIONAL ENERGY BALANCE

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8.1. INTRODUCTION

The regional energy balance exerts a major control on northern hydrological regimes as an important influence on snowmelt, in providing the energy to drive the evapotranspiration process and through the regulation of heat flow into the substrate, which ultimately controls the amount of permafrost melting in the active layer. As an example, evapotranspiration/precipitation ratios show, on an annual basis, magnitudes in the order of 0.4, 0.5 and 0.2 in the Subarctic, Low Arctic and High Arctic respectively and, for the snow-free period only, corresponding ratios of approximately 0.7, 1.0 and 0.9 (Price and Woo, 1988; Rouse, 1984c; Woo *et al.*, 1983b). Evapotranspiration thus comprises a major component in the surface water balance especially in the snow-free period. Up to 18% of the net all-wave radiation available at the surface is employed in melting of the active layer in ice-rich permafrost during the thaw period, whereas during freeze-back, this energy is released leading to a prolonged zero curtain effect (Rouse, 1984b; Halliwell and Rouse, 1987). The latter can maintain wet soils in the Low Arctic in a partially-thawed state until late December. Thus ground water in the active layer is free to move in liquid form during the early winter period. In an era of potential large-scale climate warming due to the "greenhouse effect", the interactions of higher temperatures with the energy balance will be profound. They will be felt especially through earlier snowmelt, greater available radiant energy, enhanced evapotranspiration, and changing patterns of heat exchange with the frozen substrate. The change in ground heat flux is not readily predictable because it is linked to the evaporation regime and to surface dryness.

This chapter looks at the basic processes which control the energy balance in high-latitude cold regions, the regional variations in these processes and the relationship of the energy balance to other hydrological variables. It presents an historical background to energy-balance studies in the high latitudes, a summary of current research activities in Canada and speculates on future research needs.

8.1.1. Basic Processes

The energy balance at the surface is given by:

$$Q^* = Q_E + Q_H + Q_G + Q_M \qquad [8.1]$$

in which Q^* is net all-wave radiation, Q_E is latent heat flux, Q_H is sensible heat flux, and Q_M is the latent heat of melt for snow and ice surfaces. The ground heat flux, Q_G, can be subdivided into:

$$Q_G = Q_S + Q_L \qquad [8.2]$$

with Q_S representing the storage of sensible heat and Q_L the storage of the latent heat of fusion in the ground ice.

One of the most practical uses of Equation 8.1 employs what is known as the Bowen ratio energy balance (BREB) approach to estimating Q_E whereby, for snow-free terrestrial surfaces:

$$Q_E = (Q^* - Q_G)/(1 + \beta) \qquad [8.3]$$

The Bowen ratio (β) is both a measurable term and a diagnostic term, which can describe the evapotranspiration behaviour of the surface. It is given by:

$$\beta = Q_H/Q_E = \gamma\, \Delta T_a/\Delta e \qquad [8.4]$$

in which γ is the psychrometer constant. ΔT_a and Δe are the vertical air temperature and vertical vapour-pressure gradients taken over the same height intervals (Δz) within the surface boundary layer.

Another useful expression, which has been an especially potent tool in hydrology, uses what has become known as the equilibrium model of evapotranspiration. It is expressed as:

$$Q_E = \alpha\, s/(s + \gamma)\, (Q^* - Q_G) \qquad [8.5]$$

in which s is the slope of the saturation vapour pressure - air temperature curve and α is an evaporability term (Priestly and Taylor, 1972). When $\alpha = 1$ true equilibrium evaporation exists. This represents the lower limit of potential evapotranspiration since, theoretically, it is the minimum possible evaporation from a saturated surface. Under certain conditions evapotranspiration from wet surfaces has been found to have an average value of 1.26 (Stewart and Rouse, 1976a; Marsh *et al.*, 1981; Marsh and Bigras, 1988). In the northern latitudes, however, this is probably more fortuitous than a result of describing accurately the basic physical processes, since other studies over saturated surfaces indicate that α can be highly variable (Rouse *et al.*, 1987; Bello and Smith, 1990).

The atmospheric fluxes in Equation 8.1 are described by:

$$Q_H = -\rho\, C_p K_H\, \Delta T_a/\Delta z \qquad [8.6]$$

$$Q_E = -\rho\, C_p/\gamma\, K_w\, \Delta e/\Delta z \qquad [8.7]$$

in which ρ is air density, C_p is the specific heat of air, and K_H and K_w are the turbulent transfer coefficients for sensible and latent heat. Equations 8.6 and 8.7 can be made operational in the form of aerodynamic flux equations (Halliwell and Rouse, 1989) in which the turbulent transfer coefficients are replaced by the log-linear wind profile, Von Kármán's constant and a correction for atmospheric stability. These can be useful in hydrological studies since they provide an independent means of calculating evapotranspiration which does not require the measurement of net radiation and the ground heat flux as in Equation 8.5.

A particularly useful diagnostic expression for vegetated terrain uses a resistance format to express the Bowen ratio in the form:

$$\beta = [r_c + r_a - r_i]/[(s/\gamma)r_a + r_i] \qquad [8.8]$$

where r_c is the surface resistance to evaporation and is a function of leaf stomatal resistance to water loss and the leaf area index; r_a is the aerodynamic resistance to vertical fluxes and is a function of the vigour of atmospheric mixing due to buoyant and forced convection; r_i has become known as a climatological resistance which can be expressed as:

$$r_i = (\rho\, C_p/\gamma)D/(Q^* - Q_G) \qquad [8.9]$$

where $D = e_s - e$ is the vapour-pressure deficit (VPD), e_s being saturated vapour pressure.

Equations 8.3 and 8.8 and 8.9 can be combined to give the well-known Penman-Monteith equation (Monteith, 1981):

$$Q_E = [s(Q^* - Q_G) + C_p D/r_a]/[s + \gamma(1 + r_c/r_a)] \qquad [8.10]$$

The application of Equation 8.10 to different northern surface types is discussed in Lafleur and Rouse (1988). It is evident that atmospheric control over the evapotranspiration process through the climatological resistance or vapour-pressure deficit can be as important or more important than factors of surface wetness.

8.1.2. Regional Variations

The net all-wave radiation (Q^*) can be expressed in terms of its component fluxes as:

$$Q^* = K{\downarrow}\,(1 - a) + L{\downarrow} - L{\uparrow} \qquad [8.11(a)]$$

$$= K^* + L^* \qquad [8.11(b)]$$

in which a is surface albedo, $L{\downarrow}$, $L{\uparrow}$ and L^* are the incoming, outgoing and net long-wave radiation respectively and $K{\downarrow}$ and K^* are the incoming and net solar radiation. Trend lines of $K{\downarrow}$, K^*, Q^* and L^* are plotted in Figure 8.1 using measured data from stations in Canada. In Figure 8.1, Subarctic refers to the mixed open-forest tundra lands which lie north of the main boreal forest and Arctic refers to the open tundra north of tree line. The arctic zone can be subdivided into Low, Middle and High Arctic. Low Arctic is the zone immediately poleward of tree-line which is composed of relatively lush tundra vegetation. It is confined to the Canadian mainland and the southern Arctic Islands. The Middle Arctic is a transitional zone in the Islands where tundra vegetation is less well-developed than in the Low Arctic. The High Arctic comprises the remaining non-glacierized portions of the Arctic Archipelago and is largely unvegetated.

A number of important patterns emerge in Figure 8.1. The net long-wave radiation loss shows a small but regular decline with latitude. The subarctic and arctic stations, as delineated in Figure 8.1, have average L* values which are 82 and 74% respectively, of the average for the temperate-latitude stations. The smaller L* values in higher latitudes result primarily from surface radiative temperatures which are substantially colder relative to those of the overlying atmosphere than is the case for lower latitudes. Figure 8.1 indicates that annual totals of K↓ decrease regularly with latitude. The latitudinal trends in K* and Q* indicate more pronounced decreases with latitude. Using the data from Figure 8.1 the percentage ratio of Temperate : Subarctic : Arctic for K*/K↓ is 80:73:57 and for Q*/K↓ it is 45:30:14. It is the high-albedo surfaces, which are responsible for the very small net radiation. The high-albedo surfaces result from the extreme longevity of the snowcover and its high reflectivity. The latter is ameliorated in subarctic environments by open forest and other tall vegetation (Davies, 1963; Rouse and Bello, 1983; Rouse, 1984a; Robinson and Kukla, 1985). The Arctic, suffers from both the long snowcover season and the scarcity of tall vegetation.

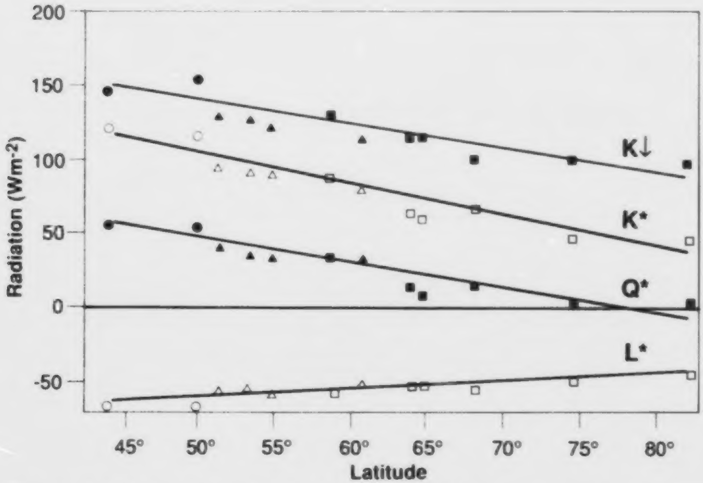

Figure 8.1. Trend lines of latitudinal variation in solar radiation at the surface, K↓, absorbed solar radiation, K*, net radiation Q* and net long-wave radiation, L*, accompanied by values for select stations in temperate ○, subarctic Δ and arctic regions ☐.

The poleward decrease in the net radiation, which is the primary forcing factor in the energy balance, means that less energy is available for Q_G, Q_E and Q_H. Regional variations in each of these are more complex than those for Q*.

The ground heat flux, Q_G, as the main energy source for melting the permafrost active layer, is an important factor in the hydrological regime.

It is, however, not well known. Estimates of its magnitude, in terms of Q_G/Q^* in spring and summer, range from 5% in dry non-permafrost subarctic environments through 12% in ice-rich wet tundra in the Low Arctic to 14% in the High Arctic. As well as responding to the magnitude of Q^*, Q_G is also influenced by air temperature and surface wetness. In terms of the vertical ground heat flux:

$$Q_G = -K_g \, \Delta T_g/\Delta z \qquad [8.12]$$

in which K_g is thermal conductivity and T_g is ground temperature. A cold air temperature favours a cold ground surface temperature which decreases ΔT_g in Equation 8.12. A dry surface promotes small values for K_g. In fact, in peat soils, the drying of the surface layers can almost completely decouple the ground thermal regime from the atmosphere above. Thus, overall drainage conditions and the vigour of Q_E are important to the magnitude of Q_G.

In the most general terms, the latent heat flux (Q_E) and the sensible heat flux (Q_H) will decrease with increasing latitude, as a response to decreasing Q^*. In the high latitudes, Bowen ratios are large compared to their temperate-latitude counterparts. Reported average values range from 0.67 and 0.50 for a subarctic dryland and wetland respectively, through 0.88 for a wet sedge grass surface in the Low Arctic, to 0.74 for upland tundra in the High Arctic. Factors other than surface moisture and available radiant energy are important to the magnitude of evaporation. Potential evapotranspiration, in the classical sense of a radiation-driven process is not an entirely valid concept in northern environments. This arises from Equation 8.8. As temperatures decrease, the vapour-pressure deficit decreases and the climatological resistance (r_i) promotes larger Bowen ratios. In terms of Equation 8.7, the vertical vapour-pressure gradient (Δe) decreases which subdues Q_E. On the other hand, cold air temperatures give larger values of ΔT_a, in Equation 8.6, and this enhances Q_H.

8.1.3. Relationship to Other Hydrological Variables

The surface energy balance is linked to all hydrological processes addressed in this book. For glacial hydrology during the ablation period, the linkages operate through the net radiation, the sensible- and the latent-heat fluxes. Typically, during ablation, a glacier surface is dirty and it is wet. This results in a decrease in surface albedo and an increased Q^*. This occurs on both a diurnal basis and as the ablation period progresses. With a warm atmosphere (0°C), Q_H is directed toward the surface and during ablation is used almost entirely in the melt process. Direct conductive heating of the glacier is only important when the surface ice is below freezing and largely ceases when surface layers reach 0°C. During ablation, Q_E may be directed upward or downward depending on the difference in vapour pressure between the saturated surface and overlying atmosphere. Negative Q_E during windy, warm, advective storms can play a major role in glacier melt, particularly in northern climates that fall under a maritime influence. Condensation and fog in the lowest atmospheric layers can serve to shield the glacier surface from the sun in the same manner as does a layer of stratus clouds. In the non-ablation period, the glacier mainly cools through long-wave radiation loss. In the high latitudes, this energy loss is normally compensated for by the downward directed Q_H during the arctic night. Q_E is negligible during this period because of the cold atmosphere.

Many of the same features which apply to glacial hydrology also apply to snow hydrology. During the ablation period the snow surface temperature does not rise above 0°C, the same as for the glacier surface. Thus Q^* and Q_H are the main processes in promoting ablation. Since many of the northern lands are in subarctic open forest, the trees play an important role in the energy balance. They serve to reduce the surface albedo thus increasing K^*, and they also reduce L^*, or in some cases, even drive it positive. Thus the forest tends to be a warmer environment than its non-forested counterpart. Because it commonly traps more snow, snowmelt from the forest is larger and more rapid than from equivalent open areas. However, because of greater snow depths, it will also commonly persist longer.

The surface energy balance is the most important forcing factor in determining the depth of permafrost, the depth of the active layer and the rate of active-layer freeze-back. Assuming that present-day permafrost is approximately in equilibrium with its environment, then it is the active layer that is the important hydrological variable. As shown in Equation 8.2, the ground heat flux either warms the mineral and organic soils, or it melts ice. In the typical "zero curtain" progression, both the soil and ice warm to 0°C, after which, Q_G is used entirely to melt ice. Only after most of it has melted, can the water and soil warm into the positive temperature range. The depth of the active layer then depends on the magnitude of Q_G and the amount of soil water. As noted earlier, if the soil surface layers remain moist, thermal conductivities will be large and Q_G will remain substantial throughout the thaw period. If, however, the surface dries out, the frozen ground becomes decoupled from the surface energy balance. The situation reverses during freeze-back when frozen ice-rich surface soils promote heat loss and subsurface freezing, whereas dry, frozen, surface soils inhibit heat loss. A low density snowcover plays the same role. The importance of the relative conductivity of surface soils is most evident in organic soils, where volumetric moisture contents or air contents can range from 0 to 90%. Thus, any experienced northerner, looking for frozen ground near the surface in summer, will go to the nearest peat hummock.

The evaporation component of the water balance is the latent heat flux (Q_E) in the energy balance. Thus, the terms evapotranspiration and latent heat flux are commonly used interchangeably. In general, as one moves poleward the relative magnitude of evapotranspiration within the water balance increases. This arises because the magnitude of precipitation declines more rapidly than that of evapotranspiration. In the short summer period, daily totals of Q_E can be of similar magnitude to temperate latitudes because of the long daylight period, which gives large Q^*, and because of the dry atmosphere which gives large vertical vapour-pressure gradients whose importance is shown in Equation 8.7. Thus typically, in the High Arctic Islands, after snowmelt run-off, all further water loss is due to evapotranspiration.

8.1.4. Importance to the Northern Environment

Although there is great variability from year-to-year, the overall hydrological regime responds to the average climate and to average climatic forcing functions. The components of the energy balance provide the main forcing functions and each of these components will be influenced by long-term climate change. The most likely imminent change is that due to "greenhouse" warming. The global models, of course, incorporate changes in the surface energy balance into their forcing functions which predict warmer and wetter conditions or warmer and drier conditions. Warmer implies a longer snow-free season. Thus snowpack melting would occur earlier in the spring and the main run-off burst would occur earlier, the annual absorbed solar radiation would increase and this would be accompanied by the potential for greater ground heating and more evapotranspiration. If the climate change results in more rainfall, there will be more evapotranspiration and summer run-off may not be enhanced. Under these conditions ground thermal conductivity will be large and deeper active layers should result. If there is less rainfall, the ground surface will become dry. Permafrost active layers may not increase rapidly, especially if a thinner snowcover leads to enhanced winter cooling of the ground. A change toward warmer and moister conditions would result in a rapid migration of the forest zone poleward so that hydrological and energy-balance characteristics of the Subarctic would be found in the Low Arctic. "Greenhouse" warming would result in earlier break-up of lake ice and the earlier disappearance of floating ice. The enhanced warming of lake waters would increase evaporation which in lake-rich regions would increase overall atmospheric humidity and possibly enhance the magnitude of local rainfall. In addition, heat loss and freezing in early winter would be retarded. Among other possible effects, this could serve to magnify warmer regional temperatures in this season, and possibly, even to increase early winter snowfall. A warmer and moister scenario would lead to greater ablation and accumulation rates in continental and alpine glaciers, whereas a warmer and drier situation would favour only more rapid ablation. In either case the short-term response would promote greater glacial run-off.

8.2. HISTORICAL BACKGROUND

Almost all energy-balance research in northern latitudes post-dates World War II (Ohmura, 1982c). For convenience these studies will be considered in the three categories, of subarctic open forest and sedge meadow, arctic tundra and glacierized arctic basins.

8.2.1. Subarctic

A number of early studies in the Subarctic concentrated on the radiation and energy balance using airborne techniques. Bryson and Kuhn (1962) overflew large portions of the Canadian Arctic and Subarctic and calculated sensible and latent-heat fluxes for a two-dimensional cross section of the atmosphere. Of necessity this early work relied on many assumptions, which resulted in Bowen ratios at the surface which are generally very small, when compared with subsequent ground-based studies (Ohmura, 1982c).

A number of detailed airborne studies provided albedo data for subarctic regions in Quebec and in central Canada (Jackson, 1960a; Davies, 1963; Kung *et al.*, 1964; McFadden and Ragotzkie, 1967). These studies provided reliable data which helped Hay (1970) and Hare and Hay (1974) to construct the regional radiation and energy balance of northern Canada.

The first major attempt to measure evapotranspiration in the Subarctic was launched by Sanderson (1949) who used irrigated lysimeters at Norman Wells. A similar study by Nebiker (1957), who experimented with lichen-covered lysimeters at Schefferville, Quebec, identified the low evapotranspiration capacity of lichen-dominated surfaces, resulting from the lichen's inability to transpire. Also working in the Schefferville area, Findlay (1969a) carefully documented the water balance of a basin which gave a derived Bowen ratio of 0.44 (Hare and Hay, 1974). About the same time, Brown (1965) measured radiation and evapotranspiration for various vegetation covers near Norman Wells. In this partial energy-balance study, he attempted to relate the existence of permafrost and the depth of the active layer to the surface energy balance with somewhat inconclusive results. The first comprehensive energy-balance calculations for the open woodland of the Subarctic were made by Pugsley (1970). For the snow-free summer period he calculated a ground heat flux (Q_G) at 5% of Q^* and a β of 0.92 (Table 8.1).

Reference	Location (Altitude) Site Characteristics	Season	Q^*	Q_G	Q_G/Q^* %	Q_E	Q_H	β
Pugsley (1970)	District of Mackenzie							
	(70 m)	melt	9.9	0.4	4	4.9	4.6	0.94
	Open woodland	summer	13.2	0.6	5	6.6	6.0	0.91
	Closed woodland	melt	11.8	0.8	7	7.2	3.8	0.52
		summer	14.1	1.5	11	6.7	5.9	0.88
Rouse *et al.* (1977)	District of Mackenzie							
	(60 m)							
	Open-lichen woodland	melt	11.4	0.5	4	7.2	3.7	0.51
	Burned-lichen woodland	summer	10.1	0.3	3	5.4	4.4	0.82
Wright (1981, 1982)	Schefferville, Que.							
	(600 m)	summer	8.5	0.3	4	5.0	3.2	0.64
	Upland tundra							
Lafleur & Rouse (1988)	Southern James Bay							
	(5 m)							
	Upland sedge	summer	9.7	0.9	9	4.8	4.0	0.83
	Sedge marsh	summer	10.0	0.9	9	5.4	3.7	0.69
	Woodland swamp	summer	10.7	1.1	10	6.1	3.5	0.58

Table 8.1. Comparative results of select surface energy balance studies in the Subarctic. Flux values are in $MJ \cdot m^{-2} \cdot d^{-1}$.

Since 1970 there have been a number of studies in the Subarctic based on field measurement. Rouse (1976) and Rouse *et al.* (1977) reported on measurements made for mature lichen woodland and burned woodland in non-permafrost areas of the western subarctic. They found that, in the mid-summer period, Q_G was about 4% of Q^* and that β was 0.51 and 0.82

for unburned and burned surfaces respectively. In a unique study at Schefferville, Nicholson (1978) snow-fenced an area of mine tailings and compared the ground temperatures and ground heat flux to an adjacent control plot over a five-year period. The snow-fenced plot had an average maximum snowcover of 97 cm whereas the control plot average 10 cm. Whereas Q_G, in summer, showed a similar gain for each plot, in winter, the heat loss exceeded the gain for the control plot, whereas the trial plot had virtually no heat loss. At the end of five years, active-layer depths in the trial plot were 2.5 times those of the control. This illustrates conclusively the importance of winter snowcover to the annual ground thermal regime.

Also working at Schefferville, Wright (1981, 1982) reported on the water balance of a lichen-heath tundra underlain by permafrost. He found that, for the snow-free period, Q_G was less than 5% of Q^* (Table 8.1). The evaporative flux over the lichen-dominated surfaces indicated values for α (in Equation 8.5) which ranged from an average 1.26 when the surface was wet after a rain to an average of 0.74 when the surface had dried. This equates to β's of approximately 0.35 and 0.95 respectively. Singh and Taillefer (1986) examined the impact of synoptic-scale advection on evapotranspiration rates from black-spruce forest in northern Quebec. With warm-air advection over a moist surface, the enhancement of evapotranspiration often exceeded 25%.

Riseborough and Burn (1988) reported on a study in an open-forest moss vegetation growing on organic soils near Mayo in the Yukon. They concluded that the moisture in the organic layer is decoupled from the mineral soil beneath, that evaporation from the organic layer declines as drying proceeds and that the mineral soils remain cold in summer because the thermal diffusivity in the organic layer is small. In quite a different study, Lafleur and Rouse (1988) studied the influence of surface cover on energy partitioning and evaporation in a subarctic wetland. They were working in a wet-sedge marsh near the south coast of James Bay in a non-permafrost area at the southern boundary of the subarctic zone. They found for this very wet area that with respect to Equation 8.10 the aerodynamic resistance (r_a) was moderate and did not vary much; the canopy resistance (r_c) was highly variable and was least for cool moist winds and greatest during warm dry ones; and that the atmosphere, operating through the vapour-pressure deficit (D) favoured large evaporation during warm and dry conditions and small evaporation for cool moist conditions. The net result for this area, under conditions of the same net radiation and a non-limiting water supply, was that β's ranged between 0.20 with a warm dry atmosphere and 1.00 with a cold moist one. It can be concluded from this, that synoptic conditions are very important and that potential evapotranspiration, in the traditional sense of a largely radiation-dominated process, does not exist. This concept will be explored in more detail later.

8.2.2. Arctic Tundra

Research in the arctic tundra has proceeded with large-scale modelling and with ground-based systematic studies. Alaska tundra is continuous with that in Canada and, since much of the Alaskan work predates and compliments the Canadian work, it will be considered in this review.

Of the larger-scale investigations the earliest comprehensive studies are found in the work of Vowinckel and Orvig (1964a, 1964b, 1970, 1971). They calculated the radiation balance for the entire polar basin, and the synoptic heat budget for three stations in the Canadian Arctic. These studies were valuable in establishing overall spatial patterns in the radiation budget and hence in the total of surface to atmosphere heat fluxes, though in detail, they suffered from the assumptions that had to be made for the many unknowns. Other synoptic energy-balance studies during the same period were reported by Vowinckel (1966) in a study at Resolute, and Barry (1972) who calculated the summer flux divergence over the Queen Elizabeth Islands.

There have been a number of important books and book chapters which deal with energy-balance calculations in the Canadian Arctic. One of the earliest of these, by Hare and Hay (1974), presents regional patterns of the energy balance and develops the difficulty of accurate determination of precipitation in calculating water balances in the tundra. Burns (1973, 1974) has produced a synthesis of the climate of the Mackenzie Valley in which some energy-balance information is presented. An important compilation by Maxwell (1980) details radiation measurements and energy-balance calculations for the Canadian Arctic Islands. Most recently, Rouse (1990) has summarized measurement-based energy-balance studies in the Canadian Subarctic and Arctic.

The earliest surface energy-balance measurements were attempted at Barrow, Alaska by Bryson (1956) and Mather and Thornthwaite (1956, 1958). The latter study covered a period of more than one year of experiments and provided an early and reliable source of information with respect to the annual energy balance of arctic tundra. Subsequently there were extensive radiation measurements made and these are summarized in the report of Maykut and Church (1973). Three important energy-balance studies dealt with the melt season at Fairbanks (Wendler, 1967) and at Barrow (Weller and Benson, 1971; Weller *et al.*, 1972). A classical year-round energy-balance synthesis at Barrow for the International Biological Programme (IBP) was presented by Weller and Holmgren (1974) and is summarized in Table 8.2. The effect of topography on the surface energy balance has been treated by Wendler (1970) and Brazel (1972) and numerical simulation studies on the surface energy balance and heat conduction in the tundra have been carried out by Outcalt *et al.* (1975) and Outcalt and Brown (1976).

The earliest surface energy-balance measurements in the Canadian tundra began in the High Arctic with radiation studies. Larsson (1963) presented results of albedo measurements over arctic surfaces. Barry (1964) gave the results of the measurement of global and net radiation at Tanquary Fiord, Ellesmere Island, Dahlgren (1974) presented results of measurements of direct and diffuse solar radiation and albedo made near sea level on Devon Island. The first energy-balance study is associated with Ahrnsbrak (1968) who measured surface fluxes at three sites in Keewatin over a 13-day period. A classical and monumental study of the surface energy balance during the late winter, snowmelt and summer seasons is associated with the work of Atsumu Ohmura on Axel Heiberg Island (Ohmura 1981, 1982a, 1982c). Begun in 1969 the work extended through 1970 and 1971 and provides the

Reference	Location (Altitude) Site Characteristics	Season	Q^*	Q_G+Q_M	$\frac{(Q_G+Q_M)}{Q^*}$ %	Q_E	Q_H	β
Weller & Holmgren (1974)	Barrow, Alaska	Premelt	1.6	1.0	63	0.3	0.3	1.00
	(7 m)	Melt	5.8	2.2	38	3.5	1.1	0.31
	Open tundra	Postmelt	15.9	1.4	9	11.6	2.9	0.25
		Summer	10.0	0.2	2	6.6	3.2	0.48
		Freeze-up	4.1	0.1	2	3.1	0.9	0.29
		Midwinter	-1.5	-0.2	13	-0.1	-1.2	12.00
Addison (1977)	Devon Island (5 m)							
	Vascular plants	3 summer	11.9	0.3	3	1.7	9.9	5.82
	Lichens	days	13.4	1.5	11	4.2	7.7	1.82
	Unvegetated		13.2	1.4	11	2.5	9.3	3.72
Addison & Bliss (1980)	King Christian Is. (20 m) Lichen-moss-rush	Summer	8.9	1.2	13	2.3	5.4	2.35
Ohmura (1982a)	Axel Heiberg Is.	Premelt	1.6	0.2	13	0.5	0.9	1.80
	(200 m)	Melt	7.2	2.4	33	2.4	2.4	1.00
	Upland tundra	Summer	7.8	1.1	14	4.0	2.7	0.68
Rouse (1984c)	Hudson Bay Lowlands 20-50 m							
	Upland tundra	Melt	2.3	0.8	35	0.8	0.7	0.88
		Summer	10.4	1.8	17	4.4	4.2	0.95
Rouse & Bello (1985)	Wet-sedge tundra	Melt	9.3	2.0	22	3.9	3.4	0.87
		Summer	12.4	1.9	15	5.4	5.1	0.94
	Inland lichen heath	Melt	10.5	3.2	30	3.9	3.4	0.87
		Summer	11.9	1.3	11	5.4	5.2	0.96

Table 8.2. Comparative results of select surface energy balance studies on arctic tundra. Flux values are in $MJ \cdot m^{-2} \cdot d^{-1}$.

most careful documentation, of the energy balance of an upland tundra dominated by hummocky sedge grass, that has been carried out in the High Arctic (Table 8.2). Ohmura found that, during the dry-snow season of late winter, net radiation was the main source of energy and the sensible heat flux into the atmosphere was the main sink. This gave a large Bowen ratio due to the coldness of the atmosphere, which creates large vertical temperature gradients and very small vapour-pressure gradients. During the melt season, the most energy was used in melting the snow. Immediately after snowmelt, in June, Q_H and Q_E increased. On average, for the snow-free period of July and August, $\beta = 0.68$, which indicates that evaporation is important in this environment, but that there is also a substantial resistance to it. During the same period, Q_G was 9% of Q^*, indicating that a substantial portion of energy was used in thawing the active layer of the permafrost. Woo (1976a) determined evaporation as part of the water balance in a study near the head of Vendom Fiord on Ellesmere Island, on a similar site to that of Ohmura's. He found that, for the high summer period, the evaporation was 150 mm, which is 60% greater than that measured by Ohmura over the same time interval. Woo noted that evaporation was the most important factor influencing changes in the suprapermafrost water table, and thus in influencing summer run-off. As part of the International Biological Programme (IBP) a group from the University of Alberta carried on energy-balance studies in the Truelove Lowland on Devon Island. This is an area whose microclimate gives a floristic diversity characteristic of much lower arctic latitudes. Working through a four-year period, Courtin and Labine (1977) measured and estimated the radiation fluxes for all seasons. They also measured soil temperature, moisture and active-layer depths, for upland beach ridge and wet-sedge meadow sites. Addison (1977) reports on measurements for three days in mid-summer at the same two sites. He determined very large Bowen ratios for sites dominated by vascular plants and by lichens and for unvegetated surfaces (Table 8.2). In a similar study over polar desert on King Christian Island, Addison and Bliss (1980) calculated an average β of 2.4 (Table 8.2). These studies give a general indication of the magnitude of site-to-site differences in the High Arctic, although the methodology and limited time span present limitations on the accuracy of the actual measurements. Smith (1975) developed a numerical simulation model of microclimate and ground thermal regimes for dry and wet sites on Ellesmere Island. Using measured and assumed surface energy-balance parameters, he was able to accurately model the soil temperature at depths of 30 and 20 cm at each site, for the period mid-June through mid-September. Using lysimetric estimations of Q_E, Marsh *et al.* (1981), working on Cornwallis Island near Resolute, found that for unvegetated upland mineral soils, α in Equation 8.5 varied between 1.26 and almost 0 depending on the wetness of the soil. This represents the most simple interface of mineral soil and atmosphere, with no intervening vegetation layer.

Energy-budget studies in the Middle and Low Arctic started later than those in the High Arctic. Initial studies were associated with a group from McMaster University working in the coastal tundra of Hudson Bay. Rouse and Stewart (1972) and Stewart and Rouse (1976a) worked on lichen-dominated upland tundra, wet-sedge tundra and shallow lakes during the mid-summer period. Average β's for those three surfaces were 0.67, 0.38 and 0.35 respectively (Rouse *et al.*, 1977). *Of some importance, these studies reported larger* β's during cold than during warm conditions. Subsequently Rouse (1984a, 1984b, 1984c) reported on radiation and energy-balance studies in the vicinity of the arctic tree line near Churchill. He found that, because of deep winter snow accumulation near tree line, snowmelt in the forest could lag behind that of the tundra by up to a month. However, at all times, winter and summer, the forest has a larger net radiation than open tundra, due primarily to smaller albedos. In addition, he found that in the wet soils of the Hudson Bay Lowlands, the heat flux into the substrate (Q_G) utilized 16 and 18% of Q^*, in tundra and forest respectively. Most of the energy involved in the soil heat exchange is used in melting the ice-rich active layer of the permafrost. During freeze-back of the permafrost, the release of latent heat is a major heat source to the surface. On dry gravel soils of the sedge-covered upland tundra, the average Bowen ratios were 0.88 and 0.95 during and after snowmelt respectively (Table 8.2). The latter value is a little larger than that found by Ohmura (1982a) for high-latitude upland tundra, and indicates a substantial resistance to evaporation. On a nearby coastal wet site, dominated by sedge grasses, Rouse and Bello (1985) reported β's of 0.87 and 0.94 for melt and post-melt seasons respectively, a result almost the same as for the dry upland site. They also got very similar results for upland lichen heath at Marantz Lake, a site 60-km inland from the Hudson Bay coast. Rouse (1984c), Rouse and Bello (1985) and Rouse *et al.* (1987) have studied extensively the impact of cold, moist summer air advection from Hudson and James Bays, on the climate of the adjacent terrestrial tundra and forested surfaces. Their studies indicate that, when winds blow onshore, which they do for about one-half of the time during May through September, Q_H is strongly enhanced and Q_G and Q_E are suppressed, compared to offshore winds. This effect extends to at least 65-km inland. Lafleur and Rouse (1988) account for the decreased evapotranspiration in terms of a climatological resistance which responds to very small vapour-pressure depressions during cold, moist onshore winds. The cold advective source is attributable to the longevity of summertime sea ice in Hudson Bay and northern James Bay. Rouse and Bello (1985) calculate that earlier melting of the sea ice, which would, for example, accompany "greenhouse" warming, would have the effect of shifting the latitude further south or of moving the tree line and boundary of continuous permafrost further northeast. This would be in addition to overall terrestrial warming.

Other studies indicate that mesoscale advection exerts a pronounced impact on temperature, humidity and the energy balance. Rannie (1977), in a summertime study of the Lewis Glacier on Baffin Island, found that, on warm days with gentle off-glacier winds, the advection of cold air caused an average temperature depression of 2.3°C at the glacier margin, and also caused a high frequency of evening inversions in the adjacent valley. Jacobs and Grondin (1988), in examining the influence of an arctic large-lakes system on mesoclimate in south central Baffin Island, conclude that the lakes act as heat reservoirs, which moderate temperatures locally during cold-air intrusion in summer, and which retard seasonal cooling in fall and early winter. Advection on a local scale also exerts a pronounced influence on the energy balance. This is documented in an important study by Bello and Smith (1990), of a shallow tundra lake in the Hudson Bay Lowlands. They show that the latent heat flux from the lake exceeds available radiant energy over the summer, with temperature inversions over the lake on the majority of days. The advective enhancement of regional advection is due both to air flow from the warmer and drier upwind terrestrial surfaces and to the synoptic effects of daily weather variations. Its impact is closely related to lake size. Marsh and Bigras (1988) in an interesting study of lake evaporation in the Mackenzie Delta similarly noted the importance of lake size. They found that for one lake in mid-summer an α of 1.26 closely described evaporation but that evaporation from another lake of different size was quite different.

Advection on a microscale can also be large. Halliwell (personal communication), on sunny days, has measured horizontal surface temperature differences over a few metres, of up to 10°C, in hummocky-sedge tundra. The wet areas are cool and the drier areas are warm. This leads to large difference in vertical soil temperature profiles in the top layers of ground over short horizontal distances. It leads also to a horizontal atmospheric flux of sensible heat, which enhances evapotranspiration from the wet areas. For wet and dry terrain types like this, which are common, it means that modelling of ground thermal regimes should account for the two-dimensional nature of the processes.

Unique features of high-latitude vegetation can exert a substantial influence on the surface energy and water balance. Roulet and Woo (1986a) report on a study carried out to the west of Baker Lake in central Keewatin, in a continuous permafrost region of wetland bogs and fens. They found that wetlands were as effective in evaporating water as were nearby lakes, due to the ability of capillary water from the water table to keep the bog surfaces wet. One of the most interesting vegetation types in the tundra is ground lichen, which is fully non-transpiring. Bello and Arama (1989) have shown that, even in tundra regions with relatively large precipitation, a lichen canopy can intercept up to 61% of the summer rainfall. This never reaches the soil and is subsequently evaporated directly into the atmosphere. The maximum storage capacities of the lichen mat are more than double those reported for forests. They suggest that a combination of open forest and ground lichen, such as is found in many subarctic regions, presents the potential for even larger overall canopy storage.

A recent summer energy-balance study in alpine tundra has been carried out by Bailey *et al.* (1989) and Bowers and Bailey (1989) for Plateau Mountain in the Canadian Rocky Mountains. Working over a surface comprising bare rock and gravel, mosses, lichens and grasses they achieved results consistent with those in low-arctic studies. As a proportion of Q^*: Q_H, Q_E and Q_G were 52, 42 and 6% respectively giving an average Bowen ratio of 1.24. The authors note that their data were gathered during a particularly dry summer.

8.2.3. Glacierized Basins

Energy-balance studies on glaciers in the Canadian Arctic have been largely limited to the ablation season and understandably have been concerned with the relationship between the surface heat budget and the magnitudes and rates of ablation.

Early research was undertaken out of McGill University in the 1960s. Andrews (1964) carried out measurements on the lower White Glacier on Axel Heiberg Island at an altitude of 208 m during July and August 1960. During this period, energy used in the subsurface heat flux (Q_G) and the latent heat of melt (Q_M) was more than double Q^*, the extra energy being provided by fluxes of Q_E and Q_H directed downward to the glacier surface (Table 8.3).

The Bowen ratio, for these downward directed fluxes was 1.6. Havens (1964) also reported on summer measurements made in a glacier accumulation area on Axel Heiberg at an altitude of 1530 m (Table 8.3). In this study, $Q_G + Q_M$ exceeded Q^* by 2.5 times with most of the extra energy being provided by downward directed Q_H, the downward directed Q_E being very small. Lister (1962) and Sagar (1962) presented results of early studies of the heat and mass balance which were carried out on the Ward Hunt Ice Shelf. Sagar (1964) reported on a study carried out in the névé zone of the Gilman Glacier on northern Ellesmere Island in which net radiation (Q^*) accounted for two-thirds of calculated melt, most of the remainder being

Reference	Location (Altitude) Site Characteristics	Season	Q^*	Q_G+Q_M	$\frac{(Q_G+Q_M)}{Q^*}$ %	Q_E	Q_H	β
Andrews (1964)	Axel Heiberg Is. (208 m) Lower White Glacier	Summer	6.6	13.9	211	-2.8	-4.5	1.61
Havens (1964)	Axel Heiberg Is. (1530 m) Upper White Glacier	Summer	1.7	4.4	259	-0.2	-2.5	12.50
Ambach (1963)	Greenland (1004 m) Ice Cap	Summer	7.9	8.6	109	1.8	-2.5	-1.39
Ambach (1976)	Greenland (1850 m) Ice Cap	Summer	0.8	1.0	125	1.2	-1.4	-1.17
Wendler & Ishikawa (1973)	Northern Alaska (200 m)	Summer						
	Bare moraine		6.1	0.5	8	2.6	3.0	1.15
	Snow		6.2	10.7	173	0.4	-4.9	-12.25
	Ice		3.7	3.6	97	2.3	-2.2	-0.96
Brazel & Marcus (1979)	Yukon, Mt. Logan (5365 m) Glacierized Plateau	Summer	-4.9	0	0	-4.4	-0.5	0.11

Table 8.3. Comparative results of select surface energy balance studies in arctic glacierized basins. Flux values are in $MJ \cdot m^{-2} \cdot d^{-1}$.

due to the downward-directed sensible heat flux (Q_H). Keeler (1964) studied the relationship of ablation to the energy balance on Sverdrup Glacier, Devon Island and Müller and Keeler (1969) using data from both the Sverdrup and White Glaciers noted that short-term comparison between measured ablation using stakes and ablation calculated from the energy balance was not good. They calculated that Q* could account for anywhere from 5% to 95% of the energy used in ablation. More recently, Alt (1978, 1979) has examined the summer synoptic conditions which control accumulation and ablation and hence the mass balance of both the Meighen and Devon Ice Caps. Although she didn't carry out energy balance measurements, her work points out the importance of the synoptic systems to ablation of the ice, whereby a persistent circulation which promotes southerly flow can result in a summer of ablation which can erase a positive mass balance of five years.

Holmgren (1971c) carried out a thorough study at an altitude of 1320 m on the Devon Ice Cap during late winter and the ablation seasons of 1962 and 1963. He got agreement between calculated energy sources and sinks within at least 20% and as close as 5%. Holmgren's results showed the following. In late winter, 1962, almost all of the energy from Q* was used in warming the glacier ice. Downward directed Q_H and upward Q_E balanced, with magnitudes comprising 39% of Q*. The two ablation seasons were very different. In 1962, downward Q_H comprised 41% of Q*, Q_E was very small and the heat sources were used fairly equally in warming and in melting the glacier ice. In 1963, Q* was small and was exceeded by downward directed Q_H. About 40% of the available energy was used in upward Q_E and most of the rest was used in warming the ice. There was no glacial melt. Holmgren's work points out the strong year-to-year variability in the energy balance of high-latitude glaciers due to the importance of the weather, particularly the frequency of warm, moist cyclonic storms. Brazel (1968) presented radiation and ablation data based on research in the St. Elias Mountains and L.D. Williams (1974) developed a computer simulation model of ablation based on the energy balance of Boas Glacier, Baffin Island and tested it on measured data. In the latter study the model response was reasonably good although substantial errors in predicted ablation were engendered.

Although not in the Arctic, there have been a number of careful studies done on the Peyto Glacier in the Canadian Rocky Mountains. Föhn (1973) obtained good agreement between calculated and measured melt during a short summer period in 1970. He calculated that the energy balance contribution to melt for $Q^* : Q_H : Q_E$ was in the ratio of 44:48:8 percent. Derikx (1975) reported on the heat balance and associated run-off from the tongue of Peyto Glacier. Goodison (1970) noted poor agreement between global solar radiation ($K\downarrow$) and ablation on Peyto. However, Munro and Young (1982) showed that the absorbed solar radiation (K^*) correlated well with measured ablation and melt. Munro (1975) found that Q^* accounted for about two-thirds of glacial melt and Q_H for one-third. Q_E was negligible. Munro and Davies (1977) have reported on the nature of the surface boundary layer over a melting glacier. It was found to be quite thin and to be strongly influenced by katabatic winds as was noted earlier by Holmgren (1971b) for Devon Island.

There have been a number of glacier energy-balance studies outside of Canada which are relevant to the Canadian scene. Two studies were carried out in Greenland by Ambach (1963, 1976) as outlined in Ohmura (1982c) at heights of 1004 m and 1850 m on the Greenland Ice Cap (Table 8.3). At the lower elevation, Ambach found that during the ablation season of 1959, $Q_G + Q_M$ was about 10% greater than Q^*, that the downward directed Q_H was small, but that there was a small Q_E flux upward giving a negative β. At the higher altitude, Q^* was very small, it was roughly equalled by Q_G, there was no ice melt, and a small downward Q_H was equalled by upward Q_E again giving a negative β. Wendler and Ishikawa (1973) carried out short-term heat-balance studies of moraine, snow and ice surfaces over an 11-day period in the summer of 1971. The studies were pursued at altitudes of about 2000 m on the McCall Glacier of the Brooks Range in northeastern Alaska near the Yukon border. As shown in Tables 8.3, whereas Q^* for moraine and ice surfaces was about the same, Q^* for snow was 40% less. The atmospheric heat fluxes over the moraine were both upward and large, giving a β of 1.2. $Q_G + Q_M$ for the snow was 70% greater than Q^*. A large downward-directed Q_H and small upward-directed Q_E gave a β of -12.3. Over ice $Q_G + Q_M$ was approximately equal to Q^*. Downward Q_H and upward Q_E were also roughly equal, giving a β of -1.0. A unique study on an arctic-alpine glacierized basin was carried out at a height of 5365 m on Mt. Logan in the Yukon by Brazel and Marcus (1979). Of all the studies presented in Tables 8.1, 8.2 and 8.3, this was the only one that showed a negative net radiation in summer. Another unique feature of this study, which was done in early summer, was that there was negligible warming or melting of the snow and ice. The main atmospheric flux was the latent heat of sublimation, whereby a large increment of frost accretion at the surface was supplied by overlying moist air masses from the Pacific Ocean.

Table 8.4, which summarizes select data for the subarctic, arctic tundra and arctic glaciers, brings out several important facts. In the subarctic and in the arctic tundra, Q* during the melt and postmelt season is large and triggers atmospheric and ground fluxes which are large. The subsurface heat flux in summer is sizeable. In energy-balance studies or in evaporation modelling, it should not be ignored. The data for $(Q_G + Q_M)$ in Table 8.4 are probably underestimates because of inherent errors in estimating Q_G (Rouse, 1984c; Halliwell and Rouse, 1987). Bowen ratios are large, even during melt and for wet surfaces in summer. In this respect, the high latitudes differ from their temperate latitude counterparts because of strong surface and atmospheric resistances to evapotranspiration. For glaciers in summer, Q* is small. Much more heat is used in melting the glacier snow and ice than is provided by net radiation, and the atmospheric fluxes are directed into the surface. Of these, Q_H is the largest giving large Bowen ratios.

Area	Season	Q^*	(Q_G+Q_M)	$(Q_G+Q_M)/Q^*$	Q_E	Q_H	β
Subarctic	Melt	10.9	0.6	6	6.1	4.2	0.69
	Summer	11.3	0.8	6	6.0	4.5	0.75
Arctic tundra	Melt	7.0	1.9	27	2.9	2.2	0.76
	Summer	10.7	1.3	12	5.1	4.3	0.84
Arctic glaciers	Summer	2.8	5.8	207	-0.5	-2.5	5.00

Table 8.4. A summary of the surface energy-balance studies presented in Tables 8.1-8.3. Studies that covered only a few days have not been included. Flux values are in $MJ \cdot m^{-2} \cdot d^{-1}$.

Energy-balance research in the Canadian North has been a response to a situation after World War II of almost total ignorance. For example, it was not known if Hudson Bay froze completely in the winter (Hare and Montgomery, 1949). Thus scientific curiosity played an important part, particularly with a group out of McGill University, who had been raised in the European tradition. Notable amongst these were Ken Hare, Fritz Müller, Svenn Orvig and Ernst Vowinckel. Many of the students of these researchers have continued working in the Canadian North. A second influence arose from the Canadian Meteorological Service (Atmospheric Environment Service) in response to demands from rapidly developing commercial air travel and from the military, concerned with defence of the northern approaches to North America. This resulted in a greatly expanded meteorological network, which has allowed regional energy balances to be developed; as seen, for example, in the work of Vowinckel and Orvig (1964a, 1964b, 1970), and Maxwell (1980). The Arctic Institute of North America has sponsored a number of important studies, especially in glacierized basins; as seen, for example, in the work of Lister (1962), Sagar (1962), Brazel (1968) and Holmgren (1971b, 1971c). Work centred in the University of

Alberta, under the direction of Larry Bliss, was part of the IBP and largely ceased with the end of this program. Other studies such as those of Smith (1975) and Rouse *et al.* (1977) took place under the auspices of the Arctic Land Use Research Program (ALUR). Studies out of Carleton University, under the leadership of M. Smith, and sponsored by the Geological Survey of Canada have been concerned with the impact of the surface energy budget on permafrost as seen in the work of Riseborough and Burn (1988). Research from the Université de Montréal directed by B. Singh, arose out of the James Bay Development Scheme and was concerned primarily with the energy balance of the subarctic forest (Singh and Taillefer, 1986). McMaster University has been a centre for energy-balance studies initiated by W.R. Rouse and M.-K. Woo and a number of students from McMaster are currently researching in the Canadian North. Most of this work has been driven by scientific curiosity. The National Hydrology Research Institute (and its predecessors) has been and is very active in pursuing and sponsoring energy-balance research as seen, for example, in the work of Munro and Young (1982) and Marsh and Bigras (1988).

8.3. CURRENT RESEARCH

In 1990 a large project known as the Canadian Wetlands Program will be centred in the Hudson Bay Lowlands and at Schefferville, Quebec. This is directed by the Canadian Institute in Atmospheric Chemistry and involves the Atmospheric Environment Service (AES), the National Aeronautical Establishment (NAE), the National Aeronautics and Space Administration (NASA) and a number of Canadian universities and other government groups. The project is designed to study carbon fluxes into the atmosphere such as CO_2 and CH_4 but in the process, components of the energy balance will be measured. The main project will take place in the James Bay Lowlands. Over a 20-day period in the summer of 1990, AES and K. King of the University of Guelph will be measuring the near-surface atmospheric fluxes CH_4, Q_H and Q_E using eddy-correlation and BREB methodology, at the same time as the NAE measures them from low-flying aircraft and NASA in higher-flying aircraft. There will be enough information to develop the complete energy balance during this period. Ancillary projects will be taking place at Churchill where W. Rouse, McMaster University, will be carrying out ground-based methane measurements as well as the thaw season energy- and water-balance calculations in 1989 and 1990, and at Schefferville where N. Roulet, York University, and T. Moore, McGill University will be measuring methane fluxes and determining the water balance in 1989, 1990 and where NAE and NASA will overfly in 1990.

Ongoing research on Baffin Island is under the direction of J. Jacobs, Memorial University. He is studying the regional energy balance in Baffin Island using automatic weather stations in conjunction with the Canadian Climate Centre. He is concerned with the mineral surface, lakes, glaciers and ice caps with the objective of modelling the impacts of climatic change.

The area around Churchill has been and remains a centre for energy-balance research. R. Bello, York University, is concerned with modelling the advection enhancement of lake evaporation, parameterizing local advection from simple models relating vegetation-canopy structure and rainfall-interception characteristics and in evaluating model results with respect to climatic change. Also presently working out of Churchill, P. Lafleur, Trent University and W. Rouse, McMaster University are attempting to operationalize the energy balance, using the resistance model in Equation 8.10. Having gathered data from the southern and central James Bay area they are concentrating on representative vegetation types in the Churchill area such as spruce and tamarack forest, sedge-grass communities, lichen-dominated surfaces and wetland terrain units in order to characterize surface resistance, aerodynamic resistance and climatological resistance. This will be used to model the evapotranspiration response to climatic change incorporating changing temperature, precipitation and vegetation cover. The models will be correlated using the water balance of existing watersheds.

In the western subarctic the National Hydrology Research Institute has initiated a multi-disciplinary study of the heat and moisture fluxes associated with wetlands in a discontinuous-permafrost environment near Fort Simpson, N.W.T. This program is directed by T.D. Prowse and is designed to aid in the evaluation of the effects of climate change on the hydrological regimes.

The only Canadian effort to use remote sensing to study the energy balance in northern Canada is under the direction of E. LeDrew at the University of Waterloo. It is aimed at the polar ocean and concerned with using remote-sensing information to develop the radiation balance and turbulent heat fluxes in order to understand atmospheric dynamics and study the synoptic evolution of the atmosphere in the polar basin.

8.4. FUTURE RESEARCH NEEDS

Energy-budget studies at a point, taken for their own sake, have or should have largely ended. Research efforts are quickly evolving into treating regional scales, with the hope that they can be integrated into global patterns. The driving force or incentive is the strong possibility of "greenhouse" warming and the large impact it would have in the high latitudes. The impact on northern hydrology could be immense. There are different needs for glacierized areas and for non-glacierized terrestrial surfaces.

It is evident from Table 8.3 that summer ablation from arctic glacierized basins involves energy input from Q^*, Q_H and Q_E and that the atmospheric fluxes often exceed the radiative inputs. Monitoring the system lends itself to remote sensing and to the use of automatic weather stations. Remote sensing can be used to approximate Q^* and to derive surface temperatures. Since most ablation occurs at above-freezing air temperatures, when the air at the glacier surface is saturated, the magnitude of the fluxes can be determined from the vertical temperature and humidity gradients between the automatic weather station and ice or snow surface using aerodynamic transfer functions derived from horizontal wind speed. Calibration of this approach would require surface-flux measurements of the type carried out by Wendler and Ishikawa (1973) and Munro and Davies (1977) to be done in conjunction with satellite derivations of radiation and of surface temperature. The other part to this approach is to document the frequency of synoptic conditions during ablation as pursued by Alt (1978, 1979) and Bradley and

England (1979). It is the synoptic climate that will change with "greenhouse" warming. This will see the more frequent incursions of warm moist air into the glacierized basins, with larger heat and moisture fluxes to the snow and ice surfaces. This allows the modelling of ablation for scenarios of changing synoptic climate.

The situation over non-glacierized land surfaces is more complex because of the variable proportioning of open water, non-vegetated and vegetated terrain, varying available water, the variable albedo and surface roughness and changing atmospheric conditions. Successful simulation modelling, which will reliably respond to scenarios of climatic change, must consider each of these variables. The type of model which can incorporate these variables takes the form of Equation 8.10. Net radiation (Q^*) as a function of solar radiation and surface albedo, will be most strongly influenced by changes in cloud cover and by the longevity of the seasonal snowcover. These variables can be modelled and actual data on both these features can be derived from satellite information. The surface albedo for snow and for arctic and subarctic terrain types is well-known and is an input variable to General Circulation Models. The ground heat flux (Q_G) is less well-known, but can be modelled successfully using Furore principles. This approach has been employed by Outcalt *et al.* (1975) in Alaska, by Smith (1975) on Ellesmere Island, and by Halliwell (1989) in the Hudson Bay Lowlands. It will be responsive to surface temperature and moisture and is thus linked to the atmospheric fluxes. The three resistances in Equation 8.8 link the surface with the atmosphere and can be modelled to simulate reality as shown in Lafleur and Rouse (1990). The surface resistance responds to the surface diffusion resistance of mineral and organic soils and the stomatal resistance or canopy resistance of plants. Where there is no vegetation, only the former operates. When there is complete plant cover, with large leaf area indices, only the latter operates. Where there are no transpiring plants, as in the case of a lichen heath, there is only canopy resistance governed by the moisture content of the lichens (Bello and Arama, 1989). The surface resistance can be characterized for typical vegetative assemblages and this requires substantial field research. It cannot be ignored in any successful simulation modelling, because it can be large even for very wet vegetated surfaces as already noted. The aerodynamic resistance to evaporation is primarily a function of wind speed and surface roughness. Climatological resistance is a function of air temperature and humidity. Wind speed, air temperature and humidity are reasonably well-documented from the existing meteorological network and surface roughness can be derived from vegetation types. The surface and atmospheric variables in Equation 8.10 can all be varied with changing temperature, humidity, and plant composition. The derived evaporative flux can be inserted into the surface water-balance equation and under scenarios of higher temperatures and more precipitation, unchanged precipitation and less precipitation, can be used to derive the basin run-off response for typical basins in the High Arctic, Low Arctic and Subarctic. This type of modelling can then be linked to remotely-sensed information on terrain characteristics to build up a larger picture. ❄

9

REGIONAL HYDROLOGY

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9.1. INTRODUCTION

Collins (1987a), in an extended metaphor, likens hydrological specialties to islands in an archipelago, each with their own language, customs, jealousies and fears. His point, of course, is that such insular attitudes ultimately stand in the way of progress in science. Nowhere is this point more relevant than in the Canadian North where the land is so large and varied, where so little is known of the region's waters or where the hydrological community is so small. If information outputs are to be maximized, common or shared approaches are essential, especially between researchers and data-collection agencies. New research results should be incorporated into data-collection programs to generate better data. This, in turn, would facilitate more rigorous testing of developed theory. A more direct or more essential synergism would be hard to envision. But why the concern about northern streamflow in the first place?

Part of that answer, no doubt, derives from the same compelling motive that drives men to climb mountains. The romantic vision of the North (Section 1.2.) continues to capture the imagination of many, including hydrologists. This is good because there are also more mundane reasons to become knowledgeable about northern waters. One important need is to ensure that the anthropogenic effects (Section 7.3.) do not cause irrevocable harm to the aquatic environment. The Canadian North is one of the last regions on earth where we have a chance to do it right the first time. Also, the High Arctic is well-suited to surface-water research; the systems are less complex than those in more temperate regions.

This chapter discusses elements of river flow in the Northwest Territories, and peripherally the Yukon. There are other subarctic and arctic regions in Canada which likely should be included in the ensuing discussions on northern streamflow, but personal experience will limit the discussion.

Research needs may be identified from perceived deficiencies in the existing body of knowledge, in this case the behaviour of northern rivers and lakes. Sections in this chapter therefore must review what is known about them in order to identify further research needs.

Research to acquire an understanding of river systems in the Canadian North is difficult because the region is so vast and possesses such great diversity in river regimes. The Northwest Territories alone are more than one-third the land mass of Canada, measuring 3,376,698 km^2 (Outcrop, 1986). Of this, 132,090 km^2 are covered by freshwater. To attain understanding of the surface waters of so large a region, it is logical to sub-divide the North into smaller units with uniform river behaviour. A short review of attempts to develop regional models in the N.W.T. is found in Section 9.2; similar attempts in the Yukon are mentioned in Section 1.4.2. A river-regime modelling discussion is concluded with a short description of the Mackenzie River flow model.

Another aspect of northern surface-water hydrology of relevance in this chapter is its link with other elements of the northern hydrological cycle. Of concern here are the anomalies which occur when northern precipitation data, mostly in the form of snow, are compared to streamflow data derived from physical measurement. These process topics are discussed in Section 9.3 in the context of four distinct elements in the annual streamflow cycle; open-water, freeze-up, midwinter, and break-up. Some of the difficulties associated with collection of streamflow data are briefly discussed in the last section (9.4). Problems stem from the inhospitable northern climate and how it affects both operational logistics and field instruments. The conclusions summarize the perceived knowledge deficiencies as they apply to both research and data collection in the N.W.T. which leads to suggestions for future avenues of research related to northern river systems.

9.2. HYDROLOGICAL REGIONS

River behaviour in large land masses such as the N.W.T. can be better understood if the larger unit can be disaggregated into smaller pieces with more uniform characteristics. Regionalization of river behaviour continues to be a topic of interest since a good characterization of regional hydrology in the N.W.T. is of great economic and scientific benefit. Some historical background for its development in the N.W.T. follows.

9.2.1. Shawinigan Engineering Studies

Early attempts to define homogeneous hydrological regions in Canada are found in the contract studies done by the Shawinigan Engineering Company for the Inland Waters Directorate (IWD), now a part of Environment Canada (DOE). For the N.W.T., two reports dated 1970 and 1982 are extant. Initial studies of hydrological regionalization by Shawinigan (1970) considered most of the N.W.T. mainland as part of a larger region which included the Prairies. The mountainous western region was considered in a Pacific/Yukon analysis. The later 1982 study was commissioned to analyse the N.W.T. by itself, and then only the mainland, since no significant hydrological data were available in 1981-82 for the Arctic Islands. Both the 1970 and 1982 studies used the same approach in which statistically significant relationships were sought between physiographic characteristics of mainland N.W.T. and annual precipitation, temperature, evaporation, run-off, flood flows and low flows. Central to the analysis was the application of the 'Square Grid Technique' to the development of regional maps for precipitation, evaporation and run-off. In this model, precipitation equalled evaporation plus run-off. Square grid land units were based on the 10 km x 10 km Universal Transverse Mercator Projection elements overprinted on the 1:250,000-scale National Topographic Series maps. Physiographic data for each square grid, as compiled in 1970, were re-used for the 1982 study. Twenty-seven variables were considered: grid location (2); elevation, slope, wind, radiation, distance-to-sea (4); barrier height (4); shield effect (4); regional slope (3); fraction of lake, forest, swamps, glaciers, and urban areas; and drainage area. Data for the same variables for river basins were taken to be the average of the values assigned to grid elements within the basin.

Multiple-regression techniques derived a first series of temperature, precipitation and run-off data for 50 calibration basins. The model temperature data were used to compute basin and grid evaporation using Turc's (1954) method of calculation. The derived values were then optimized by successive iterations of modified values until there was no further convergence between synthetic and real data. The refined model data for precipitation and run-off was then modelled against physiographic variables using multiple regression.

The final models showed that temperature could be predicted best from inputs of latitude, elevation, distance-to-the-sea to the north and distance-to-the-sea to the southwest, and, precipitation by the shield effect and distance-to-the-sea to the north. The equation for prediction of run-off contained four independent variables; elevation, latitude, shield effect and distance-to-the-sea to the north.

The physical model for run-off was based on data for 50 hydrometric stations with data from 1961 to 1980. Seventeen additional stations were used to validate the model. Error analysis of the model results generated a standard error of approximately 30%. Map plotting of the spatial distribution of error statistics indicated that several regions showed deviation from overall model homogeneity. The non-conforming regions were the Keewatin, the central Arctic and the western Cordillera.

Regional models of floods and low flows from the square-grid technique required the introduction of an additional level of complexity into the modelling system used for run-off. Statistical parameters of floods and low flows were regressed against physiographic variables at a discrimination of one month. Correlations were used to develop instantaneous flood data from the monthly models. Unfortunately, continuing high levels of uncertainty attached to the improved 1982 model made it unsuitable as a reliable water-resource planning tool.

In the last decade, the numerical data base of the square-grid inputs have been upgraded and the streamflow data base managed by IWD is being extensively reviewed. These initiatives will no doubt bring about some improvement in the application of square-grid techniques to the N.W.T. but creation of a successful model depends on the presence of a dense hydrometric network in northern Canada, something that is economically unrealistic at the present time (see Section 1.3.2.8.). The model will continue to serve a useful, optimizing function in regions with high-density data bases where down-sizing and cost cutting are the order of the day. Some of the underlying concepts could also be used subjectively for the design of hydrological networks in regions with little or no existing information.

9.2.2. Acres Consulting Services Ltd.

Acres (1976, 1982) conducted parallel studies on behalf of the Department of Indian Affairs and Northern Development (DIAND). The 1982 study was an update of previous work in 1976 and used the same assumptions and techniques. It was commissioned because of the greatly improved streamflow data base; the usable number of hydrometric stations had increased from 28 in 1976 to 51 in 1982. In 1976, they classified the N.W.T. into four relatively-homogeneous hydrological regions; the criteria being climate and physiography (Table 9.1). The new data available in 1982 were insufficient to justify further subdivision.

1. The **Cordilleran Region** - an area of the N.W.T. west from the Mackenzie River and the Franklin Mountains to the east of the river.
2. The **Interior Plains Region** - from the Mackenzie River in the west, eastwards to a line through Great Bear Lake, along Camsell River, through Great Slave Lake and southwards along Slave River.
3. The **Canadian Shield Region** - on mainland N.W.T., from the Interior Plains in the west to Hudson Bay in the east.
4. The **Arctic Islands Region**.

Table 9.1. Acres (1976) classification of N.W.T. hydrological regions.

The Arctic Islands region had almost no streamflow information, so only a very general description of the flow regime was possible. Acres' (1976) analysis includes, however, a physical depiction of the islands which divides the archipelago into upland and lowland regions and, within that broad classification, defines four regions north of the Arctic Circle based on physiographic features.

Within this framework, the studies examined algorithms for predicting annual yield, floods and low flows for the mainland regions. These employed different combinations of climatic, streamflow, terrain, geological and vegetation parameters as outlined below.

9.2.2.1. Annual Yield

Estimation of mean annual yield for an ungauged catchment on the mainland was based on the extraction of interpolated data from mapped isolines of mean run-off derived from the 1962-80 flow data for 51 hydrometric stations. Isoline placement was modified by consideration of a companion map of mean precipitation and from qualitative assessment of evapotranspiration.

Confidence limits were described from derivation of coefficients of variability (C_v), from univariate analysis of the 51 streamflow data sets and the assignment of regional coefficients. The site-specific analyses showed little regional uniformity, but generalized trends for C_V were used to assign numerical values to C_V for each region: for the Cordillera, 0.18; the Interior Plain, 0.41; and the Shield, 0.24 (upper and lower confidence limits defined by C_V were those found in Chow, 1964).

With no long-term streamflow data in the Arctic Islands, Acres (1976, 1982) proposed the use of precipitation isolines as a surrogate for flow data, arguing that precipitation and run-off were roughly equal. They suggested that the estimates of annual yield for the islands be increased from the precipitation values by a factor of 1.5 because of the underestimation of precipitation and redistribution of snow by wind.

9.2.2.2. Low Flow

Low flow estimates for 5-, 10- and 20-year return periods were made for the mainland regions only, based on algorithms derived from multiple regression of: basin area, stream length, stream slope, lake area fraction, annual mean daily air temperature, February mean temperature, annual precipitation and mean annual run-off. Regression and backwards elimination of variables, against 30-day low flows as derived from individual station low-flow frequency analysis, produced three general equations, one each for the three mainland regions.

In the Cordillera and the Interior Plains regions (station sample size 11 and 10, respectively), the low-flow regime could best be described as a power function of the mean annual run-off. In the Shield (sample size 15), the addition of two other independent variables, lake area and February temperatures, provided the best statistical fit.

The predictor equations for low flow regionally were sufficiently poor that their use beyond initial feasibility studies was not recommended.

9.2.2.3. Floods

Estimation of floods at selected return periods was based on non-dimensional regional flood-frequency curves and regional equations for the estimation of the mean annual flood. The latter were derived in similar fashion to those for low flow and used the same array of independent variables. Mean annual floods were found to vary directly as power functions of drainage area in the Cordillera; directly as power functions of drainage area, basin slope and inversely as mean annual temperature in the Interior Plains, and; directly as basin area and inversely as lake area and annual temperature in the Shield.

Acres suggested that the flood-prediction algorithms be used for preliminary feasibility studies only; they were too poorly defined to be of use in engineering or design studies.

9.2.3. Joint DOE-DIAND Studies

Since the 1975-inception of federal-provincial or territorial cost-sharing agreements for the operation of Canada's hydrometric network, its development in the N.W.T. has been funded according to approved five-year network expansion programs. The initial plan for the late-seventies had fallen behind schedule due mainly to escalating costs for building and operating new gauging sites. By 1983, the original five-year plan had been fulfilled, thus giving rise to a new five-year planning proposal. By way of justification for new stations, whose thrust was towards the definition of regional hydrology, the proposal contained a fairly detailed examination of components of the physical environment which might describe a hydrological region. Landforms and climate were two criteria that were deemed to be significant. The hydrological regions, outlined or suggested by the Acres (1976) study, were subsequently incorporated into a joint DOE and DIAND hydrometric expansion program proposal which outlined six hydrological regions (Canada, Environment Canada and Indian Affairs and Northern Development, 1983), reproduced here in Figure 9.1.

The assumption that physiographic uniformity was a principal determinant of homogeneity in river systems is evident from the station criteria shown in Table 9.2. The scanty data available in 1983 continued to be inadequate to define hydrological regions rigorously, but the analysis provided a good rationale for an enhanced network in the North.

Comparison of regional boundaries and isolines of annual yield for mainland N.W.T. in Figure 9.2, suggests that there is some merit to the proposed linkage between physiographic and hydrological regions. There is, however, need to examine these connections more rigorously with the much larger data set now available.

9.2.4. 1985-86 IWD Network Evaluation in the Western and Northern Region

Western and Northern Region (WNR) staff of IWD established a network planning committee in 1985 which subsequently developed an integrated network plan for the political jurisdictions within the region. The N.W.T. report (Wedel, 1986) identified 60 possible hydrological regions wholly or partly within the territory. These were founded, not on hydrological data, but on the assumption that combinations of landform and vegetation zones approximated regions of uniform streamflow characteristics. The choice of physiographic regions were those defined by Bostock (1967) and the National Atlas of Canada (Canada, Energy, Mines and Resources, 1973, Plate 5-6). This selection criteria was defended on the grounds that units such as plains, plateaus, mountains, hills and coastal lowlands, directly influenced the nature of streamflow. The choice of vegetation zones (Canada, Energy, Mines and Resources, 1973 - Plate 45-46) as an additional criteria was defended on the basis that it represented, in a readily-identifiable form, an integrated expression of precipitation, soil conditions, evapotranspiration and duration of the growing season - all of them useful elements when streamflow in terms of annual yield is considered.

The arguments for this detailed level of subdivision for mainland hydrological regions were also supported by the work of WNR hydrologists in the denser networks of Manitoba, Saskatchewan and Alberta. They were able to support their definitions of hydrological regions devised from the same criteria with data.

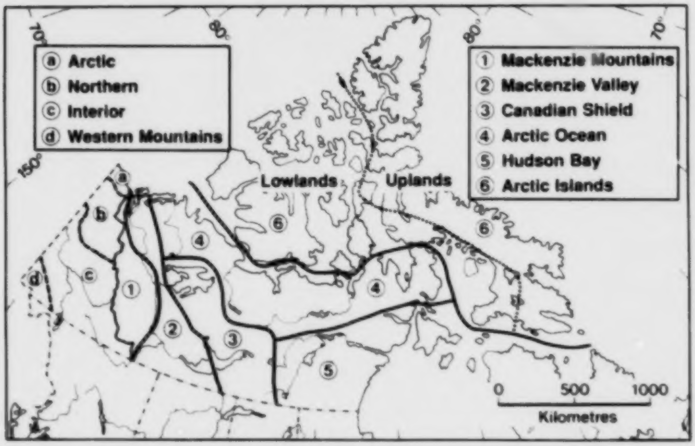
ⓐ Arctic
ⓑ Northern
ⓒ Interior
ⓓ Western Mountains
① Mackenzie Mountains
② Mackenzie Valley
③ Canadian Shield
④ Arctic Ocean
⑤ Hudson Bay
⑥ Arctic Islands
Lowlands
Uplands
70°
80°
80°
70°
150°
0
500
1000
Kilometres

Figure 9.1. Yukon and N.W.T. hydrological regions.

REGION	SUBREGION	DESCRIPTION
Cordillera	Mackenzie Mountains	mountainous area west of Mackenzie River, execept for Franklin Mountains, mainly alpine/subalpine vegetation, high precipitation producing frequent summer floods.
Inland Plains	Mackenzie Valley	low elevations in boreal forest extending east of the Mackenzie Mountains to the Canadian Shield and north to the tundra.
Canadian Shield	Canadian Shield	rugged area consisting of frequent rock outcrops, lakes and limited extent and depth of soil, forest-tundra vegetation coverage.
Canadian Shield	Arctic Ocean	drainage north of the tree line to the Arctic Ocean, flat terrain with shallow soil development in glacial tills and coastal marine deposits, uplands in the east.
Canadian Shield	Hudson Bay	drainage east to Hudson Bay, shallow soil development in glacial tills, coastal marine deposits and uplands in the north.
Arctic Islands	Lowlands/ Uplands	undifferentiated due to lack of stations, consists of upland areas on eastern Baffin and Ellesmere Islands, with arctic/alpine vegetation, glaciers, lowland areas to the west, with marine deposits along the coasts and shallow soil development.

Table 9.2. Hydrological regions of the Northwest Territories.

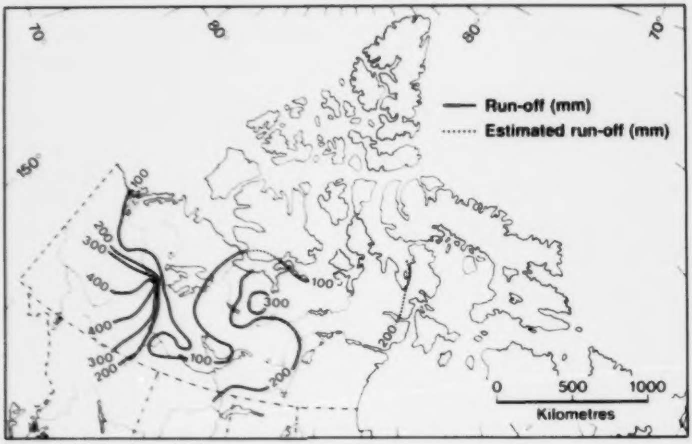

Figure 9.2. N.W.T. total run-off for 1985.

The delineation was performed on overlays for the physiographic and vegetational data on a 1:5,000,000 base map. Each physiographic unit was assigned the National Atlas number and the vegetational sub-divisions within a unit were assigned alphabetic subscripts. Such sub-regions were assigned names by association with a dominant physical feature; e.g., that part of the Kazan Upland which contained Gordon Lake became the Gordon Upland. A segment of the final map is reproduced in Figure 9.3.

Not all of the data from the 140 hydrometric stations in the N.W.T. could be used to assess the merit of this designation of hydrological regions; many of the catchments were too large to provide information for individual regions. Although some of the ones selected spanned several regions, a total of 106 hydrometric stations were ultimately assigned to regions. Only 37 of the 60 regions so identified contained any streamflow data.

The 23 remaining regions were those identified in the islands including Baffin. A few areas of the N.W.T. mainland, such as the central Arctic shown in Figure 9.3, have a density of gauging stations which makes some analysis possible. A subsequent study of annual yield between Great Bear Lake and Bathurst Inlet (Wedel *et al.*, 1988) suggests broad agreement between annual yield and the regional boundaries in the WNR network study.

While the arguments for this pragmatic delineation of hydrological regions seem to be supported in the western and central Arctic, there is no immediate hope of its being applied throughout the northern territories. Even if the 106 stations suitable for the definition of hydrological regions in the N.W.T. were uniformly distributed in space, any attempt to define 60 unique regions would clearly be ludicrous. In the N.W.T., all except four of the 106 data points are on the mainland. Although there are virtually no long-term hydrological data in the Arctic Islands, some knowledge does exist about

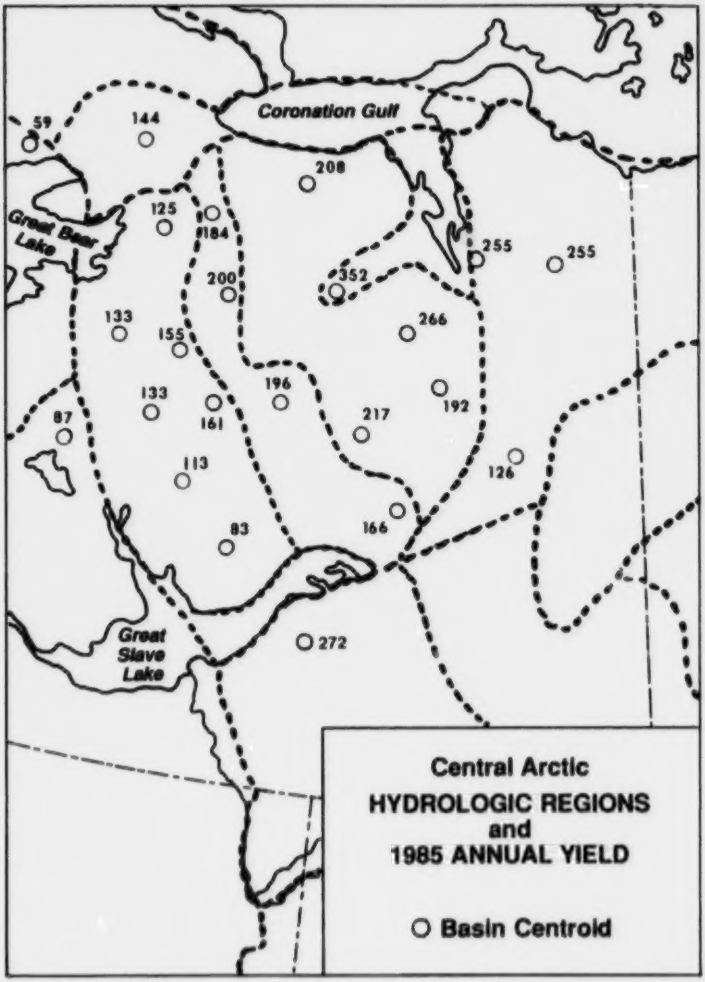
Coronation Gulf
59
144
208
Great Bear Lake
125
184
255
255
352
200
133
155
266
133
196
192
161
217
87
126
113
166
83
Great Slave Lake
272
Central Arctic
HYDROLOGIC REGIONS
and
1985 ANNUAL YIELD
○ Basin Centroid

Figure 9.3. Central Arctic hydrological regions and 1985 annual yield (mm) by basin.

streamflow behaviour in these regions. Much short-term, seasonal work has been done by the Canadian research community; see, for example, the papers or reports by Woo (1976b), McCann and Cogley (1972), Cogley and McCann (1975), Ambler (1979), Wedel *et al.* (1977a), Holecek and Vosahlo (1975), Marsh (1976, 1988), Woo (1979), and McLaren (1981). Understanding of the streamflow regime in the Arctic Islands would be enhanced by a compilation of river discharge data from the pertinent hydrological research.

Since streamflow data are so variable over time, it becomes necessary to compile them for many years in order to derive meaningful statistics on key elements of hydrological information such as annual floods, annual yield and minimum flows. Feedback from streamflow data users, compiled in the N.W.T. network analysis, showed that lack of long-term data hindered the definition of uniform hydrological regions. If 20 years of data were to be set as a minimum time series for useful analysis, then only 36 of the N.W.T. stations (Canada, Environment Canada, 1988b) and 24 of the Yukon stations would qualify. Of these, roughly half could serve as regional indicators; catchments sufficiently small to be contained in any one region.

Both the N.W.T. and Yukon networks are inadequate to address the question of regionalization for the whole of the territories. In fact, they lack the density to describe hydrological characteristics for most of the northern land mass. Some land units within the territories, however, are thought to possess networks sufficiently dense to enable development of regional parameters. These include the Central Arctic Region, centered on the Coppermine River basin, and the Plains Region in the Fort Simpson area. If models could incorporate remotely-sensed data for causal parameters like snow water-equivalents, vegetation, wetland extent, lake area, geology and physiography, and were able to combine these data with information from soil maps and from climate stations, it might be possible to develop and verify an empirical model. If rigorous theories supported the model, it could then be used to predict or describe hydrological characteristics of other regions possessing little or no streamflow information. In concept, this suggestion is similar to the square-grid model described in the Shawinigan reports (1970, 1982) but the input data could now be derived from satellite imagery. Even if the transferability aspects of models of the two regions were unachievable within acceptable limits of accuracy, their development would add substantially to the understanding of the regional behaviour of northern surface-water systems. Development of regional streamflow models, although perhaps not as appealing as other avenues of hydrological research, would be of great practical value to northern water-resource managers and would therefore be more likely to receive funding from government or industry.

It is apparent from this discussion that regional descriptions of streamflow behaviour in the N.W.T. are not yet possible with any great confidence or with any clear scientific method. Historical developments have been most successful with regionalization of annual yield. For water managers, the development of reliable, regional prediction of flood magnitude or low-flow regimes would provide them with invaluable new tools. In the absence of reliable descriptors for floods and low flows, river-routing models have demonstrated reasonably good capability to generate synthetic flow data at a daily-flow level of discrimination. From this, reasonably good estimates of flood and low flow can be made.

9.2.5. The Mackenzie River Basin Flow Model

The Mackenzie River basin was the subject of an intensive, multi-agency study in the years 1978-81 under the auspices of the Mackenzie River Basin Committee (MRBC). Its goal was "... to facilitate future cooperation in planning and management for the best use of these resources." (Letter of Transmittal, MRBC, 1981a, p.iv). Impetus for the study derived from the significant effects of Bennet Dam on the Peace-Athabasca Delta, and from the announced intentions of the B.C. and Alberta governments to develop new hydro-electric generation facilities on the Liard and Slave Rivers. Canada, Alberta, British Columbia, Saskatchewan, Yukon and the Northwest Territories participated. Nine supplements to the main report dealt with separate aspects of the overall plan (see Table 10.2). Of interest here is the modelling of the Mackenzie River by Lyons, Nemanishen and Hofer detailed in Supplement 8 (MRBC, 1981b).

The principal model utilized SIMPAK, a river flow-routing algorithm, which was designed to operate in five of the six defined segments within the Mackenzie Basin. The sixth segment, the Peace-Athabasca Delta and Lake Athabasca, utilized one-dimensional, hydrodynamic techniques. The total model terminated at Arctic Red River, just above the Mackenzie Delta. Consequently, Recommendation 6 of the MRBC summary report (Table 10.1) asked that a comprehensive study of the Mackenzie Delta be undertaken. Strategies were developed under the direction of a Mackenzie Delta Advisory Committee answerable to MRBC. A first requirement was the development of a delta flow model using one-dimensional techniques, similar to those used in the development of the Peace-Athabasca Delta segment of the Mackenzie River model. Subsequently it was to be used as the basis for a more complex, ecological model. A first phase, one-dimensional model (ONE-D) for the open-water period is now in the verification stage. Additional elements are intended to link models for sediment flux and waterfowl habitat to the fundamental streamflow model. Future development of the flow model to include winter regimes will require the participation of river-ice researchers.

Immediate uses of the Mackenzie River basin models were to assess impacts of upstream hydro-electric developments on the downstream fluvial environments. There are other potential uses, as yet undeveloped. With some modifications, the N.W.T. portion of the model could become one for flow and stage forecasting - vital to the navigational interests on the Mackenzie River and as a flood warning technique of great interest to Mackenzie River and Delta communities. One of its greatest benefits lies in the model's ability to create 'what if' scenarios of impacts of proposed developments on the Mackenzie River system. This is perceived to be absolutely essential to the whole concept of sustainable development as it relates to river systems. Can the techniques used to develop the Mackenzie model be exported to other systems within the N.W.T.? The answer is probably not, at present, because the streamflow data base in other major N.W.T. basins is of inadequate density. The SIMPAK, Peace-Athabasca and Mackenzie Delta ONE-D models were developed with data from 206+ hydrometric stations. Only 40-odd hydrometric stations are operational in the N.W.T. outside of the Mackenzie Basin.

Station Name	WSC Stn.No.	Prov.	Area km^2	Min $m^3 \cdot s^{-1}$	Max $m^3 \cdot s^{-1}$	Max/ Min
Fisher River near Dallas	05SD003	Man	1710	0.2	11.5	52.0
God's River near Shamattawa	04AD002	Man	65500	186.0	830.0	4.5
Gunisao R. at Diamond Rapids	05UA003	Man	4500	3.3	34.9	10.6
Kettle River near Gillam	05UF004	Man	1960	0.8	38.1	48.4
Little Sask. R. nr Minnedosa	05MF001	Man	2630	0.5	13.6	28.3
Manigotagan R. nr Manigotagan	05RA001	Man	1810	2.8	26.5	9.3
Red River at Emerson	05OC001	Man	104000	21.2	356.0	16.8
Brightsand River at Moberly	04GB005	Ont	1170	4.8	17.9	3.7
Magpie River at Michipicoten	02BD003	Ont	1930	10.8	83.9	7.8
Saugeen River near Port Elgin	02FC001	Ont	3960	17.4	152.0	8.7
Arctic Red River nr the mouth	10LA002	NWT	18600	10.8	519.0	48.0
Back River above Hermann R.	10RC001	NWT	93900	33.6	2170.0	64.6
Burnside River nr the mouth	10QC001	NWT	16800	7.0	495.0	70.5
Carcajou R. bel. Imperial R.	10KB001	NWT	7400	7.7	234.0	30.4
Hay River near Hay River	07OB001	NWT	47900	3.1	419.0	135.0
Quoich R. ab. St. Clair Falls	06MB001	NWT	30100	4.0	691.0	173.0
Thelon River above Beverly L.	06JC002	NWT	65600	18.4	1110.0	60.3

Table 9.3. Ratios of maximum to minimum monthly flows.

9.3. RIVER ELEMENTS IN THE NORTHERN HYDROLOGICAL CYCLE

All the gross features of southern hydrographs are found in the seasonal distribution of flow in northern rivers, but it is the details within an annual northern hydrograph that distinguishes them from their southern counterparts. For example, the snow-accumulation period is very much longer and is rarely modified by midwinter melt. Also, base-flow contributions from ground water are usually small because of the widespread occurrence of permafrost. Hence, winter flow in northern catchments tends to be much smaller than in similar-sized southern catchments. Ratios of minimum to maximum monthly mean flows for northern and southern rivers shown in Table 9.3 illustrate this point (Canada, Environment Canada, 1987a, 1987b, 1988b). This is an admittedly crude comparison since it does not consider variability in storage nor rivers whose flow drops to zero. A second feature of northern rivers is the predominance of the snowmelt flood over midsummer rainstorm floods (see Section 1.4.1. for discussion of regional hydrograph classifications). In the case of Hay River, in the northern boreal forest south from Great Slave Lake, 21 of 25 annual floods in the historical record were spring snowmelt floods. Kingyuok Creek, just outside the community of Baker Lake in the Keewatin District, has never had a recorded midsummer rainfall flood which exceeded the flood of the annual spring melt.

By far the greatest level of knowledge, in terms of scientific research and data collection, exists for the open-water season. Indeed, some researchers tend to call it the 'active season'. By implication, the other parts of the year are in the 'inactive' period when there is no surface-water flow and there are therefore no processes to study. This is true for surface-water flow in the Arctic Islands where much small catchment research has been conducted. Only two island catchments, the Union River below Stanwell Fletcher Lake on Somerset Island and the Lake Hazen drainage in northern

Ellesmere Island, are suspected of having episodic flow in the ice season following summers with above average run-off.

On the mainland, similar research on flow regimes of small basins was associated with proposals: to develop the oil and gas resources of the Mackenzie Valley and Delta, with development of the Liard and Mackenzie highways in the N.W.T., and with gas pipeline-routing proposals from the Alaska North Slope and thence along the Alaska Highway in the Yukon. Seasonal flow regimes also characterized most of these small basins.

Both logistic and scientific reasons shape the nature of the programs of the Water Survey of Canada (WSC). In the N.W.T., and to lesser extent the Yukon, cost-of-access often dictates when stations can be visited. The open-water season is when sites are most economical to visit via fixed-wing aircraft. It is also the season when by far the greatest volume of water is found in most major northern rivers. In terms of the annual yield of a river basin, little is lost by ignoring the rest of the year. Conversely, good information on minimum flow is lost, even though low-flow regimes are highly significant to northern water management.

Northern research of the last two decades has made it apparent that there exist imbalances in the data for components of the hydrological cycle. For the terrestrial segment of the cycle, gains in water from precipitation and ground water should roughly equal the losses: by surface run-off, infiltration, evaporation, sublimation and transpiration, and by changes in lake/wetland storage, especially in those large regions of the North located in the Canadian Shield.

One complicating aspect for surface-water budgets in the Arctic derives from temporal variability in the 'water year' for individual elements in the cycle. In the case of incoming precipitation, a logical starting point for the water year might be the onset of snow accumulation, since the snow accumulated over a winter determines volumes of run-off, evapotranspiration, and the gains or losses in ground and surface storage in the succeeding summer. An average date which could be considered is October 1. Site-specific refinements could be made by consideration of latitude and altitude. For run-off, the onset of the spring freshet could be considered as a feasible starting date for its water year. In large systems, or those with abundant live storage in lakes, the budget might need to consider periods beyond the following spring freshet. Evapotranspiration plays a significant role only during the growing season and sublimation of snow is a significant element of the terrestrial phase of the cycle in the six to eight week period just before and during snowmelt. These last two elements could therefore fit into the much larger time frames of the first two budget items. From this brief discussion, it becomes clear that the temporal definition of an annual water budget cannot be defined by consideration of calendar-year time frames, and further, that in large basins with a long recession period in the seasonal hydrograph, good analysis is only possible by considerations of periods greater than 12 months.

9.3.1. Snowfall Components

The complexities of water-budget analyses for large river systems have necessitated water-budget research on smaller basins. Small systems abound in the Arctic Islands and much of the surface-water information from this region derives from such analyses. Budgetary anomalies in the data were frequently evident, i.e., the data for total run-off, precipitation, evapotranspiration and storage were not in balance. Cogley and McCann (1975) pointed this out and suggested that snow was under-measured by a factor of four. Subsequent work by Woo and Marsh (1978) defined certain specifics of the shortfall in measurement of snow as it applied to a small catchment near Resolute Bay on Cornwallis Island. Wedel *et al.* (1977a) obtained similar estimates of snow under-catch from work on Bathurst Island, as did McLaren (1981), working in the vicinity of Consett Head on Melville Island in 1974. These study results likely would have exhibited even greater anomalies had they considered snow sublimation in the spring melt. A study by Holecek and Vosahlo (1975) on Truelove Lowland in northern Devon Island showed that snow sublimation in spring could account for water losses from the snowpack of up to 40 mm.

Work on the arctic mainland also suggests imbalances between the precipitation and run-off data. Way and Thorne (1978), in their study of Polar Gas pipeline crossings in the Keewatin Region, suggest that annual snowfall estimates in the vicinity of Baker Lake are low. Ongoing analysis by IWD, Yellowknife, of annual yield for the mainland N.W.T. have produced isopleths with greater values than would be expected from climatic data from weather stations in the region.

In part, the anomalies derive from dissimilarities in the data bases of Atmospheric Environment Service (AES) and IWD. Although units of measurement are the same (i.e., mm of precipitation and run-off), one essential difference derives from the point-specific nature of climatic data and the integrated average depth of run-off data for river basins at large. A further source of imbalance is the disparate temporal distribution of precipitation and run-off data, as discussed previously. Disagreement could also be attributed to the difficulties in measurement of snowfall in unsheltered, barren-land locations and to the redistribution of snow by wind during a snowfall event, all contributing to inaccurate assessment of snow precipitation.

It is suspected that better definition of snow on the ground prior to beginning of melt might significantly improve water-budget calculations in northern tundra regions. Remote-sensing techniques, in particular, offer new methods for acquiring improved snow data. The problem of distributing point-specific data in space remains a central issue for the calculation of northern water budgets. If new techniques or denser networks, or both, were to improve the reliability of spatial data, then subsequent revisions of the historical data bases could improve the understanding of northern surface-water hydrology.

9.3.2. Ice Processes and Northern Streamflow

Consideration of annual hydrographs of northern rivers with year-round flow tends to create a mental picture of two flow regimes: open-water period and ice-affected. As Davar (1979) points out, there are four regimes of which the open-water season is only one. The other three, in chronological order, are freeze-up, the midwinter period, and spring break-up, all when river ice modifies open-channel flow conditions. An understanding of the fluvial processes in these three regimes is essential to the development of good streamflow information in the ice period. Much valuable knowledge has been acquired from hydrological process research and from the WSC's data-collection activities. Short summaries of the three ice-affected streamflow regimes follow. More detailed information is contained in Section 5.3.4.

9.3.2.1. The Freeze-Up

Very little is known about northern streamflow conditions at the annual freeze-up. Inuvik staff of the WSC have obtained comprehensive flow data, often at considerable risk, in the freeze-up period on the Mackenzie River at Norman Wells and Arctic Red River. Their data show that the lower Mackenzie experiences a substantial reduction and subsequent recovery of flow in this period. The reasons for this are not at all clear. One explanation is that the modified flows are brought about by changes in hydraulic resistance during the development of an ice cover (e.g., Santeford and Alger, 1984b; see Section 5.2). A second is that flow reduction may be linked to changes in fluid viscosity produced by frazil ice crystals (e.g., Tsang, 1979, 1982). A similar phenomenon was documented by Curtis (1988) on the Diana River near Rankin Inlet in the initial stages of freeze-up during the fall of 1988. Observations which support the idea of changed viscosity under these conditions come from field staff working on ice-free rivers in sub-zero temperatures. Their accounts liken fluid behaviour to light oil rather than water. It is possible that a combination of the two - changes in resistance and fluid viscosity - might provide a better basis for understanding river regimes during the freeze-up.

9.3.2.2. Midwinter Flows

Although not as extensive as the knowledge for streamflow in the open-water season, the midwinter streamflow data base is, nevertheless, of significant size. On most rivers, the ice regime is relatively stable, and its effect on the open-water stage-discharge curve changes only gradually with time. Resistance to flow may either increase because of ice thickening or decrease in spite of ice thickening, because of decreasing roughness of the overlying ice sheet. From analysis of WSC field data, it is known that flows in many large northern rivers follow smooth recession curves (linear on semi-log paper) and, moreover, that the slope of such recession curves at a site are frequently quite similar from one year to the next.

Process research on the redistribution of velocity under rough ice cover by IWD is currently ongoing. Other recent research of interest is that by Santeford and Alger (1984a) who developed automated procedures to quantify the hydraulic effects of an ice cover on river flow. Should the procedures prove to be adequate for analysis and calculation of midwinter flows in the N.W.T., their adoption could effect significant economies in the flow computation process.

9.3.2.3. Streamflow in the Annual Break-Up

In northern Canada, it becomes difficult to collect streamflow field data in the spring break-up. Access to off-road sites in the N.W.T. and northern Yukon is restricted to helicopters. Also, physical flow measurements in periods of ice-run are either impossible or exceedingly hazardous ventures. WSC has improved their knowledge of break-up by operation of spring field camps where staff live on-site for several weeks to observe the ice run and obtain flow measurements as conditions permit. In the N.W.T., this program has been augmented by basing helicopter-equipped field crews at strategic locations such as Baker Lake, Norman Wells and Coppermine, and by increasing helicopter use from the permanent WSC bases at Fort Smith, Fort Simpson and Inuvik.

Resource constraints, both in terms of dollars and staff, make it unlikely that the IWD data-collection program will ever be able to afford more than one spring break-up camp per hydrometric station. Site specific studies of annual variability therefore, must rely on less detailed information available from the helicopter surveys.

Much information on the hydraulic effects of ice on river flow has been acquired from the spring camp program. Generally, the impact of ice backwater on open-water stage-discharge curves was found to be less than was historically used in analysis. This is especially true of the rising limb and peak of the spring hydrograph. With a few exceptions, the ice backwater effect disappears earlier in the period, often halfway up the rising limb of the freshet hydrograph. Exceptions to this generalization have been noted at the Mackenzie and Peel River stations immediately upstream of the Mackenzie Delta. Because of low hydraulic gradients, the impacts of increased water levels from ice jams as far away as the lower delta continue to affect the stage-discharge relation for days after all vestiges of river ice at the station have gone.

To date, WSC has operated field camps on Back River above Hermann River (WSC Station No.10RC001), Kirchhoffer River near Coral Harbour (06PA001), Baillie River near the mouth (10RA002), Coppermine River above Bloody Falls (10PC003), Kendall River below Dismal Lakes (10PC001) Atitok Creek near Dismal Lakes (10PC002), Ellice River near the mouth (10QD001), Thelon River above Beverly Lake (06JC002), Lorillard River above Chesterfield Inlet (06OA002), Mackenzie River at Arctic Red River (10LC014), and Peel River above Fort McPherson (10MC002). The Inland Waters Directorate (Canada, Environment Canada, 1984b) has published a report on the Back, Baillie and Kirchhoffer spring camps operated in 1983 and McDonald (1987) has prepared a report on the 1985 spring camp at Bloody Falls on the Coppermine River.

The expanded N.W.T. program of streamflow data collection in spring freshet could support process studies related to ice jamming if co-ordination between ice research and the streamflow data programs could be realized.

WSC Stn.No.	Name of Station	First Year	Latitude (N)	Longitude (W)
07QA001	Taltson River outlet Tsu Lake	1930	60°38′59″	111°56′49″
07QC003	Thoa R. outlet Hill Island Lake	1968	60°30′18″	109°38′56″
07RD001	Lockhart R. outl. Artillery Lake	1944	62°53′48″	108°28′17″
07SA002	Snare River below Ghost River	1947	63°58′30″	115°26′02″
07SA005	Wecho River outlet Inglis Lake	1983	63°10′37″	115°11′33″
07SB010	Cameron River below Reid Lake	1975	62°59′40″	113°17′05″
07SB013	Baker Cr. outl. Lower Martin L.	1983	62°30′46″	114°24′13″
07SC002	Waldron River near mouth	1978	63°02′53″	110°29′15″
07TB001	Emile River outlet Basler Lake	1945	63°47′46″	116°07′21″
09AA003	Lindeman Creek near Bennett	1954	59°50′16″	135°00′50″
09AA014	Fantail R. outlet Fantail Lake	1956	59°35′40″	134°23′26″
09AB009	Yukon River above Frank Creek	1951	61°26′04″	135°11′18″
09AC004	Takhini River outlet Kusawa Lake	1952	60°36′46″	136°07′26″
10BE004	Toad River above Nonda Creek	1961	58°51′20″	125°22′50″
10BE007	Trout R. at km 783.7 Alaska Hwy	1970	59°20′07″	125°56′11″
10HC005	Blackwater R. outl. Blackwater L.	1987	63°55′41″	124°03′56″
10JA002	Camsell River outlet Clut Lake	1933	65°36′24″	117°45′56″
10JA004	Acasta R. ab. Little Crapeau L.	1980	64°52′30″	116°08′19″
10JC003	Great Bear R. outl. Great Bear L.	1961	65°08′05″	123°31′06″
10PB001	Coppermine R. outl. Point Lake	1964	65°24′48″	114°00′14″

Table 9.4. Yukon and N.W.T. stations with no backwater from ice.

Floating ice research is described in detail in Chapter 5. Work done by R. Gerard (University of Alberta) on the Yukon and Hay Rivers, and by T.D. Prowse (National Hydrology Research Institute; NHRI) on the Liard River at Fort Simpson has provided much new information on the dynamic nature of streamflow in the unsteady flow conditions of the spring break-up. This knowledge has enabled WSC to improve their computational procedures for data production.

The presence of ice in the river reach at a hydrometric station introduces large uncertainties in determination of flow. Lake outflow reaches, however, are frequently free of ice. It therefore becomes possible to obtain highly reliable discharge data year round from hydrometric stations located in or above ice-free lake outlets. Thirteen such sites in the N.W.T. and seven in the Yukon have been identified (Table 9.4). Recently, an additional, ice-free station on the Yellowknife River has been added to the network. There are also other hydrometric stations in the N.W.T., like Back River station north from Baker Lake, where ice effect through the winter is minimal. Collectively, these stations provide the most reliable winter-flow data in northern Canada. For all other northern hydrometric stations, great uncertainties exist for winter streamflow data, especially in the periods of break-up and freeze-up.

9.4. NORTHERN STREAMFLOW DATA COLLECTION

Streamflow data acquisition and computational procedures are standardized in Canada and the United States. IWD has prepared a series of manuals which describe preferred Canadian practices (Terzi, 1981; Halliday and Terzi, 1983; Ozga, 1980). American counterparts are found in Rantz (1982). While standard operating procedures for any data-acquisition program are essential, they do create difficulties in remote regions where access is by small aircraft and the climate is not conducive to reliable operation of instruments.

The development of a continuous record of streamflow requires the definition of a stage-discharge curve which relates water flow to water level. Water-level data, which constitute the independent variable, need to be collected at each hydrometric station on a continuous basis. Periodic flow measurements are made to define the initial relationship and, thereafter, to track its shifts. Continuous stage record is then used to develop a record of streamflow. Complications arise from effects of riverine processes which alter the fundamental relationship; in the North, the most common is from the effects of an ice cover which increases hydraulic resistance to flow in a river reach. Such ice effects are dynamic, especially in the freeze-up and break-up periods when unsteady flow conditions prevail. Periodic field measurements of flow are necessary to define ice effect on the stage-discharge curve. Geomorphological changes in a river reach, common in sand or gravel bed rivers, must also be tracked by field measurement.

Flow measurements in periods of ice cover, which in northern Canada are often longer than eight months in a year, are normally obtained with instruments significantly modified from the original designs. Both the United States Geological Survey and IWD are testing improved flow meters with plastic rotor assemblies of Lexan which promise to lessen or eliminate slush-ice fouling problems. Preliminary results show a general increase of 10% in flow volume with the new equipment.

Under-ice distribution of velocities are significantly different from those in an ice-free water column. Pelletier (1988), following an extensive review of stream-gauging literature, suggests that indiscriminate use of the IWD standard methods of sampling can lead to sizeable error. In the Yukon, Alford and Carmack (1987) and Carmack and Alford (1986) have investigated this problem and suggest that IWD procedures be modified on large northern rivers. Inuvik staff of WSC have done research on vertical flow distribution at Norman Wells and Arctic Red River on the Mackenzie River. Their preliminary findings suggest that use of standard IWD methods under rough surface-ice conditions may introduce errors in the range of ±10%. New techniques capable of resolving these difficulties would be welcomed by northern hydrologists in the data programs.

At a few selected sites in Canada, some success has been experienced with acoustic instruments, but these are expensive, not portable, nor designed for use in the rigorous climate of the arctic winter. Other instruments which sense velocity in flowing water rely on measuring distortions in magnetic fields, but existing models would likely require extensive modification to make them reliable in northern working conditions.

All recent work has indicated that standardized field procedures for measurement of velocity under rough ice cover need to be re-evaluated. Research directed towards improved velocity sensing under an ice cover would not only lead to the development of better instruments, but would also improve current knowledge of the hydraulic behaviour of rivers in winter. Pratte (1979) provides an excellent summary of the then current state of research. Aside from the work in the Yukon by Alford and Carmack (1987) and on the Mackenzie by WSC Inuvik, the last decade has not seen much new research in northern Canadian rivers on this topic and the data base for field observation of distribution of velocity in the vertical is still minimal.

Measurement of stage at all hydrometric stations in the N.W.T. is accomplished by means of a mercury manometer pressure sensor linked to an analog chart recorder. In the N.W.T. a program has recently been completed to equip 102 of 140 sites with Data Collection Platforms (DCP) to provide real-time access to station water levels via GOES (Geostationary Operational Environmental Satellite) retransmissions. Unfortunately, the system requires line linkage to sensors laid in the river bed; probably the weakest link in the instrument package, since the line is vulnerable to damage from ice scour at the river bank. A method to sense water level from a shelter on shore with no physical link to the river would make IWD's river-stage sensing devices much less susceptible to failure at many northern sites.

Research and development in this small niche of hydrology has been hampered by limited markets and high costs. Successful development of new flow measurement systems will therefore likely need the support of government. In seeking this support, it should be emphasized that information about low winter flows on northern rivers is vital to the development and management of hydro-electric power, municipal water supplies and river diversions.

9.5. CONCLUSIONS

This chapter has traced the evolution of numerical models to define regional hydrological parameters in the N.W.T. and to enable impact assessment. Needs for new or modified models are identified: for regionalization in the N.W.T., for flow forecasting on the Mackenzie River as an aid to navigation, and for flood warning.

The Shawinigan (1982) and Acres (1982) regional models would probably be much improved by updates to incorporate 10 years of new data. The Shawinigan model as applied to the N.W.T. used variables from a generalized, national model. It is possible that additional variables, more specific to the N.W.T., might improve model performance. Improvement is certain to result from the ongoing IWD data review; significant changes have been made, especially to the data for annual floods. Research is needed to improve the performance of all N.W.T. models so that water-resource managers can use them beyond preliminary feasibility studies.

New studies are also required to verify the concept of many more hydrological regions than the four proposed by Acres. Consideration could be given to an increase of the hydrometric network in selected regions such as the central Arctic to test, and almost certainly to modify, the regions proposed for the N.W.T. in the IWD-WNR network evaluations of 1986. New hydrometric stations to accomplish this could be low-cost, short-term 'rover' stations which would flesh out the knowledge obtained from the existing, long-term 'master' stations.

Further developments, improvements and extensions to the Mackenzie River flow model have assumed a new urgency with recent announcements of intentions to proceed with oil/gas pipelines from the Beaufort Sea and the Mackenzie Delta. Not enough is known about sediment flux in the system. Since hydrocarbon pollutants travel on wash load particles, it becomes urgent to model their source and ultimate fate. The impact of various scenarios of development could then be easily simulated. Development of the flow model for operation in winter will certainly require the participation of floating-ice research specialists. Improved understanding of streamflow is needed for the break-up and freeze-up periods of the water year. The northern freeze-up data base is virtually non-existent, which, in turn, hinders scientific understanding of some of the underlying processes.

Finally, a better understanding of open-channel flow in the Arctic Islands would derive from the compilation of a streamflow data base from the historical research literature.

Data-collection programs of the Water Resources Branch in IWD and process research on aspects of open-channel flow in the Canadian North are linked in many ways. The research community in this discipline is not large and the data collection program is minimal. Close liaison between the two sections is essential. ✻

10

WATER MANAGEMENT NORTH OF 60°N: THE ADMINISTRATION OF INLAND WATERS

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10.1. INTRODUCTION

In 1966, the **Government Organisation Act** formed the Department of Indian Affairs and Northern Development (DIAND) and assigned its Minister responsibility for administering those resources of the Yukon and Northwest Territories (N.W.T.) not already assigned to other departments or agencies of the Canadian Government.

Between 1966 and 1968, the main thrust of DIAND was the development of the oil, gas and mineral potential of Canada's North. By 1968, however, it became apparent that a comprehensive program for the management of the renewable resources was required. Although several pieces of legislation, including the **Canada Shipping Act**, the **Navigable Waters Protection Act** and the **Fisheries Act**, were applied to the water resource none provided for specific planning and management of water. Thus, in 1970, the federal government passed the **Northern Inland Waters Act (NIWA)** (Canada, 1985) and assigned responsibility for it to the Minister of DIAND (Naysmith, 1973).

10.2. LEGISLATIVE FRAMEWORK

In 1972, the NIWA was proclaimed in force. During presentation of the Bill to Parliament, the Minister of DIAND stated that it had four purposes:

1. to provide for the equitable distribution and sharing of rights to use water in the North among interests with legitimate and sometimes conflicting claims on this resource;
2. to ensure that the disposition or allocation of water rights is done in a manner that is consistent with immediate and long-term regional and national interests;
3. to ensure that all works and undertakings planned for the use, diversion, storage or treatment of water are designed and constructed to acceptable engineering standards; and

4. to establish and maintain the principle that rights to the use of water are dependent on the users accepting full responsibility for maintaining its quality or restoring its quality to acceptable standards before returning the water to the natural environment.

Under the NIWA, the property in and the right to the use and flow of all waters are for all purposes vested in Her Majesty in Right of Canada. Exceptions to this are certain riparian rights for domestic uses by persons owning or occupying lands adjacent to waters and for emergency fire-fighting or flood-control purposes. Otherwise, no person may use water or alter or divert the flow of water unless pursuant to a water licence or as authorized by the Regulations. A rather unique feature of NIWA is that it not only applies to water use, but also to deposits of waste. A strict prohibition is placed on the deposit of waste into water unless pursuant to a licence or as authorized by the Regulations. To carry out the function of licensing water use and waste disposal, the Act established two Water Boards, one in the Yukon and one in the N.W.T.

10.3. WATER BOARDS

The objectives of the Water Boards are to provide for the conservation, development and utilization of the water resources of the Yukon and the N.W.T. in a manner that will provide the optimum benefit for all Canadians and for the residents of the North in particular. The Boards apply these objects when carrying out their responsibilities for issuing licences, with the approval of the Minister of DIAND, and for conducting public hearings on all licence applications. In addition, the Boards may make recommendations to the Minister on matters relating to these objects. Board membership includes one nominee from the federal departments of Health and Welfare, Environment and DIAND and at least three nominees from the respective territorial governments.

10.4. LICENSING AND AUTHORIZATION

10.4.1. Licensing

With the exceptions noted earlier, all water uses require either a licence, or authorization through the Regulations. If a proposed water use does not meet the criteria for authorization, then it must be licensed. An applicant for a licence must provide the Board with sufficient information to allow the Board to evaluate any qualitative and quantitative effects of the proposed use. When a Board is satisfied that it has received sufficient information, then it must call a public hearing. A hearing must be called for all licence applications, with the exception of emergency amendments. If a hearing is scheduled and no intervenors come forward by the 10th day preceding the hearing date, then the Board may cancel the hearing. At a public hearing, a Board has all the powers of a commissioner appointed under the Inquiries Act. In essence, this means that a Board can compel people to appear before it and give evidence under oath. As well, the Board may request any and all documents which may be relevant.

Once a Board has heard and considered all the information concerning an application, it may issue a licence, with the approval of the Minister, for a term of up to 25 years. In practice, licence terms have generally been for less than 10 years, with the exception of hydro-power developments. A Board may also amend, renew or cancel a licence.

In setting licence conditions, the Boards have a great degree of discretion. The Act states that "... a Board may attach any conditions that it considers appropriate".

Water-quality standards may be prescribed by Regulations upon which the Boards could base effluent conditions. However, no such standards have yet been put in place. In setting conditions, a Board must also ensure that a licence does not contravene water-quality requirements which have been set under the **Canada Water Act** or the **Fisheries Act**.

Before issuing a licence, a Board must make four very important determinations:

1. the proposed use of waters will not adversely affect the use of waters by an existing licensee or applicant who is entitled to precedence by Regulation;
2. appropriate compensation has been or will be paid by the applicant to licensees of lower priority who may be adversely affected by the proposed use;
3. any waste that may be deposited is consistent with the maintenance of prescribed water quality standards; and
4. the applicant is financially responsible.

The first of these requirements is the cornerstone of water allocation, i.e. priority of use. While the Act allows the setting of priorities in the Regulations, no such priorities have been set. This is due to the difficulties inherent in deciding which uses will have precedence over others. As well, there has been no overwhelming need. Thus, it could be argued that all uses now have the same priority and so existing licensees have precedence. The question of compensation for lower-priority licensees then becomes a moot point, although the Boards may require security from an applicant for the protection of owners and occupiers of property, and other licensees, who may be adversely affected by the issuance of a licence.

10.4.2. Authorization

The Act also allows the use of water if authorized by the Regulations. When the Act came into force, it became clear that the majority of water uses were of a "minor" nature and that the licensing process could be unduly time consuming. Thus, DIAND provided authority in the Regulations for the authorization of specified water uses by a Controller of Water Rights, one in the Yukon and one in the N.W.T. The Controllers, who were employees of the department, attached terms and conditions to the water uses which were determined by type, quantity and duration of use. As it turned out, over 90% of water uses were being authorized. For example, during the five-year period prior to 1981, the yearly average for authorizations issued was 440, while the average for licenses was 7. In 1981, the

Dene Nation and the Métis Association of the Northwest Territories filed a claim with the Federal Court challenging the legality of the authorization process. In 1984, the Court ruled that this process was in fact *ultra vires* the Act on the grounds that the Act foresees the Regulations authorizing water uses and thus, the imposition of a Controller was beyond the intent of the legislation. In order to protect the existing authorized users, DIAND amended the Regulations to conform with the Court ruling. The Regulations now simply state that water may be used without a licence for:

(a) municipal purposes by an unincorporated settlement or a construction camp;

(b) water engineering purposes; or

(c) any purpose not referred to in (a) or (b) at a rate not in excess of 50,000 gallons per day (227,305 litres).

10.5. ENFORCEMENT

The enforcement of the Act rests with DIAND. The Minister appoints inspectors who are granted authority to enter any project for inspection and enforcement purposes. Failure to comply with the Act, Regulations or a licence may result in cancellation of the licence or a fine of up to $5,000 a day for each day that an offence is committed. To date, successful prosecution has been obtained on 39 separate charges.

10.6. DATA COLLECTION

10.6.1. Water Quantity

DIAND maintains a formal agreement with the Water Survey of Canada, Environment Canada, for the operation of a hydrometric network, on a cost-shared basis, in the Yukon and the Northwest Territories. Presently, 74 stations in the Yukon and 134 stations in the Northwest Territories are maintained through the hydrometric agreement. In addition to the DIAND/DOE cost-shared network in the Yukon, DIAND maintains a hydrometric network on its own to collect data which does not fall within the objectives of the DIAND/DOE agreement. Originally, this network, which consists of 12 stations, was prompted by the lack of hydrometric information on smaller drainage basins. The acceleration of development proposals, requests for information from private interests and special interests, such as placer mining, have shaped the geographical extent of this network.

10.6.2. Water Quality

Currently, water-quality monitoring is carried out in the Yukon and Northwest Territories to address major existing resource development and water planning and management issues. The sampling frequency varies depending on the type of monitoring program required and the time frame available for collection of information. Some baseline and effects-monitoring sampling are also carried out.

All water licences issued to industrial and municipal operations require water-quality monitoring programs in response to any wastes that may be produced. As part of DIAND's enforcement functions, inspections may include the collection of additional water-quality data to supplement and sometimes audit, data collected by the licensees. Presently, water-quality data are being collected at over 70 sites in the territories.

Additional baseline and effects-monitoring data are required to enable DIAND to develop appropriate regulatory requirements and to anticipate and predict impacts of development projects. Towards those ends, DIAND, Environment Canada and the two territorial governments have recently negotiated cost-shared water-quality monitoring programs for the territories.

10.6.3. Precipitation

DIAND maintains precipitation networks in each Territory to supplement data collected by the Atmospheric Environment Service of Environment Canada.

In the Yukon, DIAND operates 12 dedicated precipitation stations and 58 snow-measurement stations, including 4 automated stations with snow pillows, precipitation gauges and thermographs. In the Northwest Territories, DIAND operates 50 snow measurement stations in the Mackenzie River basin.

10.7. PLANNING

To provide a framework for the Boards' licensing activities, DIAND is responsible for water-resource planning. The legislative framework of the NIWA allows the department to conduct comprehensive, inter-jurisdictional planning as well as regional planning, within each territory.

10.7.1. Comprehensive Planning

During the 1970s, major development proposals, both within the territories and within the provinces, pointed to the need for co-operative arrangements amongst jurisdictions which shared water resources. Since most of the major development proposal projects were within the Mackenzie River basin, that area was given top priority. Thus, in 1978, the Governments of Canada, Yukon, Northwest Territories, British Columbia, Alberta and Saskatchewan agreed to conduct a three-year, $1.6 million study of the water and related resources of the Mackenzie River Basin.

The aim of the study was to increase understanding of the resources and their interdependencies in order to better predict and appreciate the implications of water-resource development in the Basin. The study program focused on six major components:

1. Pipeline Documents Review,
2. Information Exchange,
3. Hydrological Systems,
4. Potentially Sensitive Areas,

5. Spring Flood, and
6. Public Information.

The study was completed on schedule and in August 1981 the results were made available in the form of a Study Report and nine supplemental volumes (Mackenzie River Basin Committee (MRBC), 1981a). Nine recommendations were made to and accepted by all the participating governments. These recommendations, along with a listing of the supplemental reports, may be found in Table 10.1 and Table 10.2 respectively.

The main issue identified was the need for co-operation among all the jurisdictions in managing water. Thus, the major recommendation of the report was that an agreement be reached through which transboundary water-management issues could be addressed. In response to this recommendation, DIAND initiated discussions with all parties, and negotiations are underway with all jurisdictions, with the aim of reaching an overall master agreement and several bilateral transboundary water-management agreements.

An initiative in the Yukon similar to the Mackenzie River Basin Study resulted in a 1980 agreement amongst the governments of Canada, British Columbia and Yukon to conduct a three-year Yukon River Basin Study at a cost not to exceed $2.2 million. The study was completed in September 1984 (Yukon River Basin Committee, 1984). The Yukon study dealt with nine main programs:

1. Hydrology,
2. Water Quality,
3. Fisheries,
4. Wildlife,
5. Placer Mining,
6. Tourism, Parks and Recreation,
7. Energy,
8. Socio-Economic, and
9. Information Exchange.

The objective of the Yukon study was studies which could be used to formulate a planning framework under which potential development alternatives in the Basin could be evaluated. A complete list of the study recommendations may be found in Table 10.3. Not surprisingly, the main concern identified by the study was the need for co-ordination among the agencies involved in water-resource management in the basin. Thus, the major recommendation focused on the requirement to formally establish a committee of agencies with responsibilities for water management to co-ordinate ongoing planning and management activities. Such a committee, the Yukon River Basin Committee, has been formed with representatives from the federal departments of Indian Affairs and Northern Development and Environment, the Government of British Columbia and the Government of Yukon.

It is recommended that:

1. the jurisdictions at an early date conclude an agreement through which transboundary water management issues such as minimum flows, flow regulation, and water quality can be addressed at jurisdictional boundary crossing points in the Mackenzie River Basin, and which establishes a permanent board to implement the provisions of the agreement;
2. the proposed integrated long-term network of hydrometric, snow, meteorologic, sediment, and water quality stations outlined in this report be implemented;
3. precipitation, streamflow, and lake levels be evaluated for the entire basin to determine variability over time and space, and that the relationships among all parts of the water budget be analyzed and modelled;
4. the computer model of the river system developed in this study be continually refined and updated as additional hydrometric data are collected and as changes occur on the river system;
5. the monitoring and analysis of the processes involved in spring break-up and sediment transport in selected streams in the Liard River subbasin be continued until several high-flow years have been documented, and that the study area be extended to include the Mackenzie River from Fort Simpson to its mouth;
6. a comprehensive study of the Mackenzie Delta be undertaken to determine the relationships among the processes of delta change; and the vegetation, fish, and wildlife resources; and the hydrologic regime;
7. each jurisdiction in the Mackenzie River Basin, when considering a project proposal, follow an environmental assessment process that accommodates the interests in other jurisdictions in an equitable manner;
8. the program of abstracting documents regarding water and related resources in the Mackenzie River Basin for entry in the national water document database (WATDOC) be continued;
9. the jurisdictions recognize the increasing level of interest in the resources of the Mackenzie River Basin and ensure that any new information on the resources and their development is made available to each other and to the public in a timely fashion.

Table 10.1. Mackenzie River Basin Study (MRBC, 1981a): recommendations.

1. Sensitive Areas: Literature Review WATDOC References
2. Alluvial Ecosystems
3. Spring Breakup
4. Hydrometeorologic Network Design
5. Hydrometeorologic Network Design: Background Reports
6. Slave River Delta
7. Athabasca Sand Dunes in Saskatchewan
8. Daily Hydrologic Model User Manual
9. Water Quality (1985 publication date)

Table 10.2. Mackenzie River Basin Study (MRBC, 1981a): list of supplemental reports.

10.7.2. Integrated Planning

In the late-1970s, it became apparent that the composite value of resources, including social, cultural, economic and physical factors must all be taken into account in resource decisions. In addition, a framework was required through which long-term, multi-resource planning could take place. Thus, DIAND proposed the idea of land-use planning and in 1981 approval was received for the development and implementation of a land-use planning process for the North. This task turned out to be much more difficult than anticipated but formal land-use planning agreements now exist in both the Yukon and N.W.T.

The aim of the process is to establish an orderly approach to resource-use decision making whereby local and regional participation is maximized. Through the gathering of information, long-term comprehensive plans will be produced to provide guidance for resource management, to help identify economic opportunities, to identify sensitive and unique areas for special management, and to improve consistency in decision making.

The process itself operates through the establishment of regional-planning commissions which are responsible for developing regional plans. The Minister of DIAND and the respective territorial ministers are responsible for approving the plans in consultation with the aboriginal groups.

To date, four planning commissions have been established in the N.W.T.: Lancaster Sound; Beaufort-Mackenzie Delta; Denendeh; and Nunavut. In the Yukon, a commission has been established for the Greater Kluane Region. The first complete land-use plan (Lancaster Sound) is expected to be presented to the Minister of DIAND and his territorial counterpart very soon.

While land-use planning does include consideration of inland waters, the impact of the planning process on water management remains unclear, since no land-use plans have been completed and implemented yet.

It is recommended that:

1. the governments of Canada, British Columbia and Yukon enter into a formal agreement that will establish a coordinating committee of agencies with water management responsibilities to complete the development of a framework for water resource planning and to coordinate ongoing planning and management activities;
2. government agencies coordinate water and land use planning activities. Specifically, closer ties should be established to ensure proper management of shoreline development control, access to lakes and rivers, and developments on floodplains;
3. ongoing activities for planning and management of water and related resources in the Basin should include public participation from their inception;
4. Appropriate jurisdictions review the various recommendations found within the study reports and identify, for implementation, those which can contribute to water management;
5. the following water resource items receive priority consideration and that appropriate agencies:
    - a) establish water quantity stations required to provide baseline, site specific and small stream data;
    - b) establish a water quality network;
    - c) integrate the water quantity and water quality networks where practical;
    - d) evaluate the effect of the construction and operation of existing hydroelectric facilities on the water resource;
    - e) maintain and improve the daily flow model, particularly in terms of the White River contributions and headwater lakes elevations;
    - f) evaluate the status of groundwater use and quality, not included in the study, with particular emphasis on drinking water supplies; and, where appropriate, establish land use controls;
    - h) continue examination of the cause of ice jam floods at Dawson to understand processes; and
    - i) emphasize consideration of the effects of regulation on biological productivity in the current limnological research program on the headwater lakes.

Table 10.3. Yukon River Basin Study: recommendations.

10.8. MAJOR MANAGEMENT CONCERNS

10.8.1. Mine Abandonment

Traditionally, mining has been the largest private-sector employer in the North. Currently, approximately 13 mines operate producing zinc, lead, gold and silver. All of these mines have one thing in common. They produce tailings which can include toxic chemicals, compounds and metals which were used in or produced by the milling process. These tailings are usually discharged to a retention pond where some form of treatment, be it natural or induced, is applied to render the tailings less harmful to the environment.

During the operation of the mine, the water licence addresses the problems inherent in tailings ponds such as seepage and discharge requirements. What happens though when a mine is permanently abandoned? What conditions will ensure that the long-term water quality downstream of the abandoned facility will not be degraded? How do we ensure that sufficient money will be available to implement a satisfactory abandonment plan?

The Department and the Water Boards continue to address these questions. The Water Boards now include abandonment provisions in all licences and abandonment plans are a necessity. Last year, the N.W.T. Water Board, in conjunction with DIAND, developed guidelines for mine abandonment (N.W.T. Water Board and DIAND, May, 1989). The Yukon Water Board held a public meeting this year specifically to address the question of financing long-term abandonment. In 1986, the Department released "The Northern Mineral Policy" which included the commitment to review regulatory standards consistent with good mining practices such as mine-site rehabilitation and to incorporate such standards under mining legislation (Canada, 1986). As well, the Department intends to develop a mine-abandonment policy within the next year. Clearly though, the question of mine abandonment will be an issue for years to come.

10.8.2. Transboundary Issues

A significant amount of water in the North is shared with other jurisdictions. The Mackenzie River basin in particular is shared by both territories and the provinces of British Columbia, Alberta and Saskatchewan (see Figure 1.4). While the N.W.T. contains nearly half the basin, it is downstream from all the other jurisdictions and therefore is subject to the upstream uses. The effects of the Bennett dam on the Peace River system, coupled with Alberta's interest in hydro-electric development on the Slave River pointed out the need for some form of protection from upstream impacts. The recent announcements by Alberta of increasing pulp-mill activities on the Peace-Athabasca system have again brought the matter to the forefront.

The Yukon also shares water resources with the N.W.T. and British Columbia. Although the Yukon is more of an "upstream" jurisdiction than the N.W.T., concerns over possible diversions, hydro-electric developments and other industrial activities and their impacts on the shared resource are real.

As mentioned earlier, the Mackenzie River Basin study and the Yukon River Basin study recognized inter-jurisdictional co-operation as the key to responsible water management. DIAND responded to these initiatives in an aggressive manner and is now actively negotiating transboundary water-management agreements with all shared jurisdictions. The aim of these agreements is to provide for the equitable sharing of both water quantity and quality.

10.9. NEED FOR LEGISLATIVE AMENDMENTS

The NIWA has been in force since 1972 without substantive changes having been made. Through the administration of the Act, certain deficiencies and ambiguities have become apparent. As well, changes have become necessary simply to keep up with evolving situations. Finally, the Federal Court ruling of 1984 left the Department in a position that was, while not unmanageable, certainly not very satisfactory from a water-management perspective. The need for amendments has been stated by all client groups in the North.

In response to all these factors, the Department is in the midst of preparing a comprehensive amendment package scheduled for presentation to Parliament. Basically, the amendments fall into four main categories:

1. **legal certainty** - in response to questions raised as to the rights of unlicensed water users including those which are termed "instream" users as well as certain ambiguities in the present Act;
2. **executive authority** - as a result of the ambiguity of the Act with respect to the authority of the Minister vis-à-vis the Water Boards;
3. **protection of users** - to clarify compensation and security provisions relating to all users.
4. **inspection and enforcement** - to strengthen the powers of inspectors and to bring the enforcement actions in line with current practice.

A summary of the major proposed amendments may be found in Table 10.4.

10.10. ABORIGINAL LAND CLAIMS

The comprehensive claims policy of the federal government recognizes an aboriginal interest in renewable resources (Canada, 1987). Through the negotiation of comprehensive claims with four aboriginal groups in the North, the nature of these interests will be defined, as will changes in the water-management regime.

In 1984, the federal government and the Committee for Original Peoples' Entitlement signed the Inuvialuit Final Agreement (IFA), a comprehensive claims settlement for the Western Arctic (Canada, 1984). With respect to water, the Inuvialuit have ownership of the beds of all lakes, rivers and water bodies on Inuvialuit lands, which encompass 90,650 km^2. Ownership of the water itself remains vested in the Crown. The Crown also retained the right to manage and control waters and waterways for the purpose of the management of fish and migratory birds and for the purpose of carrying out governmental functions relating to navigation, transportation, flood

1. Provide for a two-tiered water licensing system, the major elements of which are:
   a) to retain the present licensing system, with the amendments outlined below for major water use licences and create a new, more expedient licensing system for minor and non-controversial water use licences;
   b) to allow an officer designated by the Minister of DIAND to sign the "minor" licences following development of the terms and conditions by the Water Boards;
   c) to allow the prescription in regulations of the criteria to be used in determining whether a particular application constitutes a major or minor water use application; and
   d) to waive the mandatory public hearing requirement for "minor" licences but leave the Water Boards with the power to require one if the public interest dictates.
2. Explicitly provide the Minister of DIAND with the ability to issue binding directions to the Water Boards.
3. Provide the Minister of DIAND with the authority to delegate his responsibility for the Water Boards to the appropriate territorial minister.
4. Explicitly recognize domestic and instream water uses and the rights of such water users by binding the Water Boards to take into account these uses when considering the issuance of licences to other users, and by providing for the payment of compensation where a domestic or instream use is adversely impacted by another water use.
5. Enable licence conditions to include legally enforceable site specific effluent quality requirements notwithstanding the presence or absence of water quality standards as set by Regulations.
6. Clarify and strengthen the provisions of the Act which deal with compensation and security by:
   a) making both licensees and non-licensed water users eligible for compensation on provable loss or damage caused by another water use; and
   b) explicitly stating that security deposits are to be used as performance guarantees and in the absence of satisfactory performance the security may be used by the Crown to carry out the work necessary to ensure a safe site.
7. Provide for the issuance by inspectors of "stop work" orders and "clean-up" orders where there is a danger to persons or the environment and an appeal process for such orders.
8. Make it an offence to violate any provision of a licence, to fail to report a deposit of waste, or to fail to comply with a "stop work" or "clean-up" order.
9. Raise the maximum fine for an offence:
   a) for "major" licensed users from five thousand dollars ($5,000) to one hundred thousand dollars ($100,000) per day; and
   b) for "minor" licensed users from five thousand dollars ($5,000) to fifteen thousand dollars ($15,000) per day.
10. Include powers to apply fines and/or imprisonment to corporate executives found guilty of polluting the environment.

Table 10.4. Northern Inland Waters Act: major proposed amendments.

control and similar matters. Consultations must be carried out with the Inuvialuit Land Administration on these matters.

In essence, the IFA foresees water-rights allocations being carried out through the existing government process. However, any water use that may impact the bed or banks of a water body requires the consent of the Inuvialuit. Participation by the Inuvialuit was also negotiated for several planning and management bodies which impact on water management. These include land-use planning commissions, environmental-impact screening and review boards and wildlife- and fisheries-management committees.

Agreements in Principle for claims settlement have been completed, or are nearing completion, for three other native groups: the Dene Nation and the Métis Association for the central Arctic region; the Tungavik Federation of Nunavut for the eastern Arctic; and, the Council of Yukon Indians for the Yukon. All of these claims can be expected to deal with rights regarding water use, both within and outside the settlement areas, and with water and other resource-management processes, including guaranteed participation and decision-making powers. While the impacts of these claims on water management are, as yet, unclear, one thing is certain: water management in the North will be substantially altered.

10.11. TRANSFER OF RESPONSIBILITIES TO TERRITORIAL GOVERNMENTS

In 1987, the Government of Canada confirmed its commitment to the political evolution of the North by approving a political and economic policy framework to guide federal activities north of 60°. This framework was developed in response to pressures during the mid-1980s to settle land claims, divide the N.W.T., devolve government responsibilities, prepare for major development projects and provide more economic opportunities for native northerners. The result was a framework which provides a consistent and overall context for the government's future activities in the North.

One of the four objectives on which the framework is based is the transfer to the territories of provincial-type responsibilities still held by the federal government. The management of water resources is one such program. The transfer of programs has already begun including forestry, electrical utilities and health services. With respect to water, the transfer process will be driven by northerners when they consider themselves ready. As mentioned in the section dealing with amendments to the NIWA, perhaps the first step could be the delegation of authority for the Water Boards to the appropriate territorial minister. ❄

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APPENDIX I. SYMBOLS

A	coefficient in Equation 2.3
B	coefficient in Equation 2.3
A_i	icing area ($100\ m^2$)
C_i	specific heat of ice ($J \cdot kg \cdot {}^\circ C^{-1}$)
C_p	specific heat of air ($J \cdot kg \cdot {}^\circ C^{-1}$)
D	vapour-pressure deficit (Pa)
D_s	snowpack depth (m)
E	evaporation (mm)
I	net influx to the saturated layer ($m \cdot s^{-1}$)
K_g	thermal conductivity ($W \cdot m^{-1} \cdot {}^\circ C^{-1}$)
K_H	turbulent transfer coefficient for sensible heat ($m^2 \cdot s^{-1}$)
K_w	turbulent transfer coefficient for latent heat ($m^2 \cdot s^{-1}$)
$K\downarrow$	incoming solar radiation surface ($W \cdot m^{-2}$)
K^*	net solar radiation ($W \cdot m^{-2}$)
L	length of the saturated layer (m)
L_f	latent heat of fusion ($J \cdot kg^{-1}$)
$L\downarrow$	incoming long-wave terrestrial radiation ($W \cdot m^{-2}$)
$L\uparrow$	outgoing long-wave terrestrial radiation ($W \cdot m^{-2}$)
L^*	net long-wave radiation ($W \cdot m^{-2}$)
P	precipitation (mm)
Q	run-off (mm)
Q_E	latent heat flux ($W \cdot m^{-2}$)
Q_G	ground heat flux ($W \cdot m^{-2}$)
Q_g	ground-water discharge ($m^3 \cdot s^{-1}$)
Q_H	sensible heat flux ($W \cdot m^{-2}$)
Q_L	storage of the latent heat of fusion in ground ice ($W \cdot m^{-2}$)
Q_M	the latent heat of melt for snow and ice ($W \cdot m^{-2}$)
Q_R	flux of heat from rain ($W \cdot m^{-2}$)
Q_p	run-off from rainfall ($m^3 \cdot s^{-1}$)
Q_S	storage of sensible heat ($W \cdot m^{-2}$)
Q_s	snowmelt run-off from ice-free areas ($m^3 \cdot s^{-1}$)

Q_t	total discharge ($m^3 \cdot s^{-1}$)
Q_w	ground-water discharge rate ($m^3 \cdot d^{-1}$)
Q^*	net all-wave radiation ($W \cdot m^{-2}$)
R/P	run-off ratio
S_i	irreducible water saturation
S_w	water saturation
S^*	effective water saturation
ΔS	change in storage (mm)
T_a	air temperature (°C)
$\Delta T_a/\Delta z$	vertical gradient of air temperature ($°C \cdot m^{-1}$)
T_g	ground temperature (°C)
T_s	snow temperature (°C)
U	volume flux of water ($m^3 \cdot m^{-2} \cdot s^{-1}$)
U+	volume flux of water above shock front ($m^3 \cdot m^{-2} \cdot s^{-1}$)
U-	volume flux of water below shock front ($m^3 \cdot m^{-2} \cdot s^{-1}$)
V_i	ice volume [of icing] (100 m^3)
a	surface albedo
c	speed of the water particles in a saturated layer at the base of the snowcover ($m \cdot s^{-1}$)
d	grain diameter (mm)
d_o	initial grain diameter (mm)
e	vapour pressure (Pa)
$\Delta e/\Delta z$	vertical gradients of vapour pressure ($Pa \cdot m^{-1}$)
e_s	saturated vapour pressure (Pa)
g	acceleration due to gravity ($m \cdot s^{-2}$)
h	thickness of the saturated layer (m)
p_w	water pressure (Pa)
q	lateral flow in the saturated layer ($m^3 \cdot s^{-1}$)
q_L	discharge per unit width of the saturated layer of length L, at time t_L ($m^3 \cdot s^{-1}$)
r_a	aerodynamic resistance to vertical fluxes ($s \cdot m^{-1}$)
r_c	surface resistance to evaporation ($s \cdot m^{-1}$)
r_i	climatological resistance ($s \cdot m^{-1}$)
s	slope of the saturation vapour pressure - air temperature curve

t	time (s; d in [2.3])
t_i	freezing period or icing season (s x 10^{-5})
t_s	lag time due to lateral saturated flow at the base of the pack (s)
t_u	lag time due to the vertical unsaturated percolation (s)
du/dt	rate of change of internal snowpack energy ($W \cdot m^{-2}$)
x	coordinate along base of the snowpack (m)
z	elevation above a datum (m)
Λ	coefficient in Equation 2.7 ($m \cdot s^{-1}$)
α	evaporability term
β	Bowen ratio
γ	psychrometer constant
ε	coefficient in Equation 2.7
$d\zeta/dt$	rate of propagation of shock front ($m \cdot s^{-1}$)
θ	ground slope
κ	permeability (m^2)
κ_e	equivalent permeability (m^2)
κ_s	saturated permeability (m^2)
μ_w	dynamic viscosity of water ($N \cdot s \cdot m^{-2}$)
$d\xi/dt$	downward movement of a horizontal, one-dimensional wetting front ($m \cdot s^{-1}$)
ρ	air density ($kg \cdot m^{-3}$)
ρ_d	dry snow density ($kg \cdot m^{-3}$)
ρ_i	ice density ($kg \cdot m^{-3}$)
ρ_w	water density ($kg \cdot m^{-3}$)
ϕ	snow porosity
ϕ^*	effective porosity

APPENDIX II. ABBREVIATIONS AND ACRONYMS

AES	Atmospheric Environment Service, DOE
ALUR	Arctic Land Use Research Program
ARC	Alberta Research Council
B.C.	British Columbia
BREB	Bowen ratio energy balance
CRREL	Cold Regions Research and Engineering Laboratory, U.S.A.
C_v	Coefficient of variability
DCP	Data collection platform
DDT	dichloro-diphenyl-trichloroethane
DIAND	Department of Indian Affairs and Northern Development
DO	dissolved oxygen
DOE	Environment Canada
EM	electro-magnetic methods
EMR	Energy, Mines and Resources
EOS	Earth Observing System
GOES	Geostationary Operational Environmental Satellite
GPR	ground-probing radar
IASC	International Arctic Science Committee
IBP	International Biological Programme
IFA	Inuvialuit Final Agreement
IWD	Inland Waters Directorate, DOE
MRBC	Mackenzie River Basin Committee
NAE	National Aeronautical Establishment, U.K.
NASA	National Aeronautics and Space Administration, U.S.A.
NFR	non-filterable residues
NHRI	National Hydrology Research Institute, IWD, DOE
NIWA	Northern Inland Waters Act
NWRI	National Water Research Institute, DOE, Burlington, Ontario
N.W.T.	Northwest Territories

RIM	River Ice Management Program
SC	specific conductance
SEBJ	Société d'énergie de la Baie James, Montréal, Quebec
SRP	soluble reactive phosphorus
TDR	time-domain reflectometry
TDS	total dissolved solids
TN	total nitrogen
TP	total phosphorus
WNR	Western and Northern Region, IWD, DOE
WQB	Water Quality Branch, IWD, DOE
WRB	Water Resources Branch, IWD, DOE
WSC	Water Survey of Canada, WRB, IWD, DOE
Y.T.	Yukon Territory